CRREL REPORT 97-9

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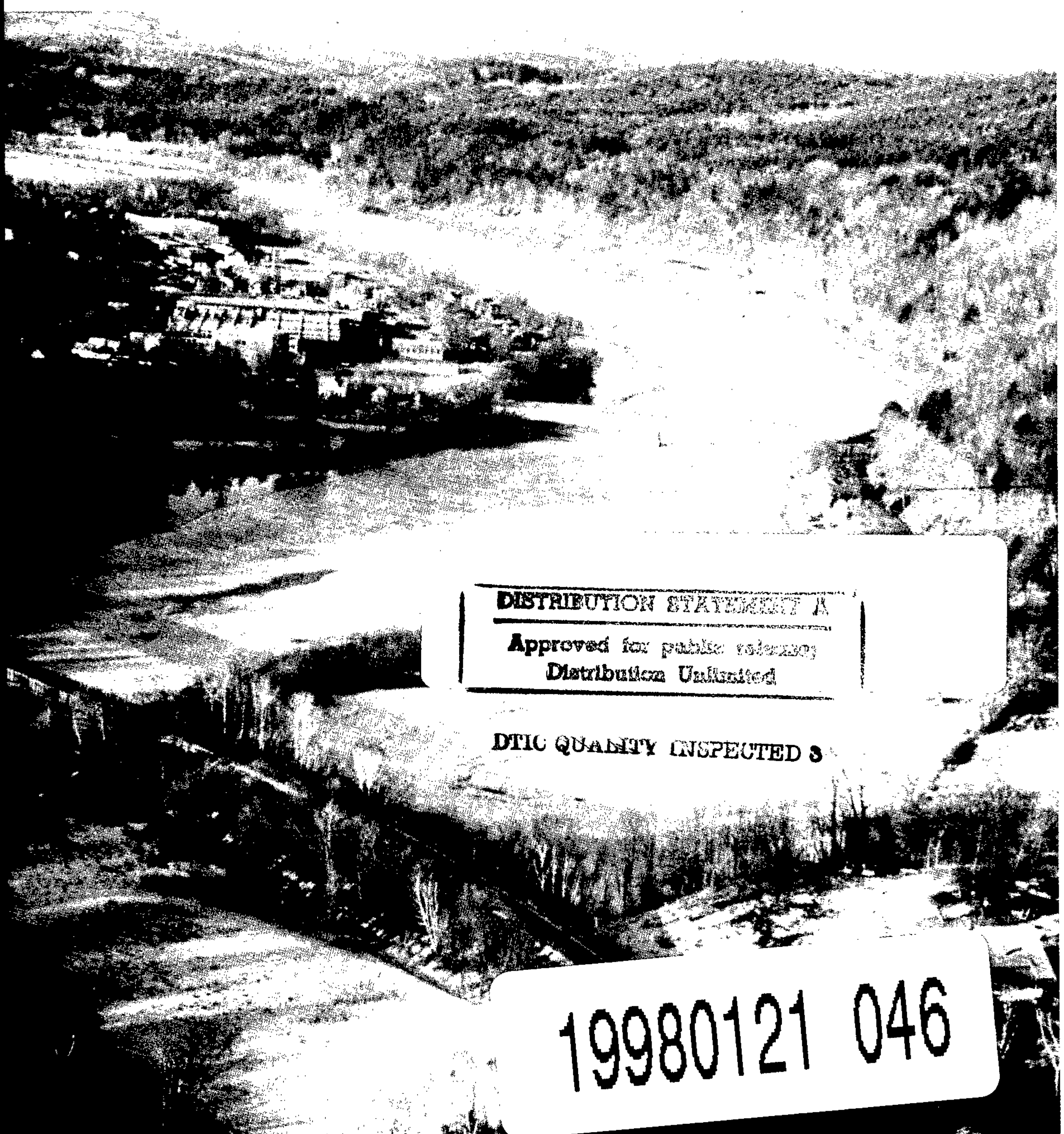
Local Variation in Winter Morning Air Temperature

Austin W. Hogan and Michael G. Ferrick

December 1997



Abstract: Results of temperature measurements, which may be applied to inference of winter temperatures in data-sparse areas, are presented. The morning air temperatures during three winters were measured at 80 places in a 10- × 30-km area along the Connecticut River. NOAA climatologies show this region to have complex spatial variation in mean minimum temperature. Frequency analysis techniques were applied to evaluate the differences in daily local temperature. Temperature lapse or temperature inversion in the study area was inferred from the difference of surface temperature measurements 100 and 300 m above river level. The frequency of inferred temperature lapse and the inferred lapse rate diminished as snow cover increased. The frequency of inferred temperature inversion and inversion strength increased as snow cover increased. When more than 20 cm of snow covered the ground, an additional surface inversion was frequent in the layer less than 100 m above river level, and two-thirds of river level temperatures less than −20°C occurred concurrent with these conditions. The daily temperature differences at the individual points, with respect to a defined point, were lognormally distributed. The magnitude and geometric standard deviation of temperature differences throughout the study area were larger on mornings when inversion was inferred. With respect to topography, temperature differences and the geometric standard deviation of temperature differences were smaller along flats or among basins than along or atop slopes on mornings when inversion was inferred. It is proposed that some meteorologically prudent inferences of surface temperature and near-surface temperature lapse or temperature inversion can be made for similar data-sparse areas.

Cover: Northward view of the Connecticut River Valley at Windsor, Vermont, and Cornish, New Hampshire (just south of the study reach) in late winter. (Photo by R. Demars.)

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Engineering Laboratory

Local Variation in Winter Morning Air Temperature

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December 1997

Prepared for
OFFICE OF THE CHIEF OF ENGINEERS

PREFACE

This report was prepared by Dr. Austin W. Hogan (retired), Research Physical Scientist, Geochemical Sciences Division, and Michael G. Ferrick, Hydrologist, of the Geological Sciences Division, Research and Engineering Directorate, U.S. Army Cold Regions Research and Engineering Laboratory (CRREL), Hanover, New Hampshire.

Funding for this work was provided by DA Project 4A161102BT25, Work Package 304, *Biogeochemical Processes*, Work Unit EC-B03, *Air-Snow-Ice-Soil Contaminant Interactions in Cold Regions.*

The authors thank Dr. Edgar Andreas and Dr. Charles Ryerson of CRREL for technically reviewing the manuscript of this report.

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NOMENCLATURE

C_p	specific heat of air at constant pressure, 1,002 J/kg°C at –23°C
$H_2–H_1$	change in enthalpy of an isolated system (J)
h_s	snow thickness on the ice (m)
H_{ia}	ice to air heat transfer coefficient (W/m^2°C)
h_{ii}, h_{if}	initial, final ice thickness (m) corresponding to times t_i, t_f
k_i	thermal conductivity of ice (W/m°C)
L	latent heat of fusion for water (J/kg)
$M = \rho V$	mass of air in the system (kg)
$_1Q_2$	heat transfer to an isolated system between states 1 and 2 (J)
T_m	melting temperature of ice (0°C)
T_a	air temperature (°C)
$T_2–T_1$	change in air temperature of an isolated system between states 1 and 2 (°C)
t	time
V	volume of air in the system (m$^{3)}$
ρ_i	density of ice (900 kg/m$^{3)}$
ρ	density of air, 1.39 kg/m^3 at –23°C and a pressure of 1000 mb (100 kPa)

EXECUTIVE SUMMARY

It is often necessary to infer the air temperature at a place where there is no regular meteorological observation. Forensic meteorologists are frequently called upon to provide an opinion as to the possibility of snow or ice covering the pavement at a place with an exposure far different from the nearest meteorological station. Military meteorologists must infer weather conditions in areas where there are few or no observations to support snowmelt analysis for flood forecasting. There are problems in engineering meteorology that require knowledge of frequency and elevation of local inversions. This report describes the influence of local topography and snow cover on observed winter morning surface air temperature. It proposes some generalizations that may be useful in inferring local air temperature when few meteorological parameters are known.

This summary introduces and summarizes a three-part report that describes the observation and measurement techniques used to obtain a daily field of 100 winter morning surface temperatures, the analytical methods applied to the data field, and several results from this research. The three parts are quite detailed in order to support some fundamental findings resulting from this work, to permit comparison of this work with that of others, and to facilitate extension of the methods used here in solution of other winter problems. The major results of this work have been published in the *Proceedings of the Eastern Snow Conference* from 1990 to 1993 (Hogan and Ferrick 1990, 1991, 1992, and 1993), and have been published in the *Journal of Applied Meteorology* (Hogan and Ferrick 1997 and in press).

Several subsynoptic-scale meteorological problems became apparent to one author (A. Hogan) during his tenure as Chief of CRREL's Geophysical Sciences Branch:

- Problems in interpretation of infrared images and apparent infrared temperatures of cold surfaces, which were obscured by air temperature inversions a few meters above the surface.
- Visual, infrared, and microwave mirages produced by air density gradients related to temperature inversions over snow.
- Extreme variability in atmospheric dispersion of contaminants and obscurants over snow-covered ground, and the difficulties involved in now-casting local temperature structure from even "nearby" synoptic observational data to provide input to dispersion models.
- Variability of winter morning surface temperatures within a single air mass and the related problem of now-casting, or even post-casting, the surface temperature at a point of interest from available data.
- Problems associated with interpreting winter climatological records, when observing sites have been relocated, or when surroundings were modified.
- The problems associated with now-casting inversion extent, and near-term forecasting of inversion persistence to provide guidance in advisories of imminent glazing.

All of these problems evolve from the spatial and temporal modification of air temperature structure over a cold surface, and the modification of the nature of the ground surface by snow cover. We hypothesized that a large, homogeneous, and quasi-level surface, as that of the Connecticut River above Wilder Dam, can provide a reference for comparison of other surface air temperatures at different eleva-

tions above other terrain. We then developed a method of atmospheric air temperature structure analysis, based on comparison of other surface temperatures, with those above and along the Connecticut River. Our major conclusions are that both the frequency of lapse or inversion, and the lapse rate or inversion strength, are related to the depth and uniformity of snow cover.

Part I describes the instrumentation, observation techniques, and analytical methods used. A moving probe technique was used to measure temperature 1.25 m above the ground from a moving vehicle. Experiments showed this technique integrated air temperature over a 60- to 100-m path and provided essentially exact repetition of measured temperature when direction of passage was reversed under lapse conditions.

Measurements of surface temperature at two elevations (100 and 300 m above river level) were used to infer lapse or inversion structure. The air temperature over the river was used as a reference, and temperature differences were extracted to form a data field. We found the logarithms of the temperature differences to be normally distributed with respect to number of observations. The geometric standard deviation of temperature differences was greatest, in general, on days that inversion structure was inferred, and specifically greatest at stations that coincided with the bases of inversions.

Temperature differences were stratified with respect to snow depth, temperature structure, and distance along the river from the reference point. There was less than 1.0°C difference among points within 10 km of the reference point and less than 1.6°C difference at 30 km from reference point on 84% of trials.

We conclude that this verifies our hypothesis, and indicates that atmospheric conditions influencing sky radiation and turbulent exchange of heat to the surface were homogeneous along a 30-km north–south path during our experiments.

The temperature differences at many elevations within 6 km of the river were stratified by snow cover and inferred vertical structure in Part II. This stratification indicated that lapse rate diminished with increasing depth and uniformity of snow cover. Inversion strength increased with increasing depth and uniformity of snow cover. Lapse structure was most frequent below in the layer 100 m above river level when less than 20 cm of snow covered the ground. There was most frequently a surface-based inversion below the 100-m inversion when more than 20 cm of snow was present.

Temperature differences of 6°C to 18°C occurred over distances of 0.6 to 2 km west of the river on days when less than 2°C of temperature difference was measured over distances of 30 km along the river. We conclude that terrain differences rather than sky cover differences are responsible for these large temperature variations, which generally coincide with local inversions a few tens of meters thick. Some repetitive measurements, made on two days when temperatures less than –39°C were observed, showed that the temperature structure typical of most winter days was preserved.

Part III relates the results of the first two parts to some of the problems listed. Comparison of temperature measured in a hamlet to that measured in surrounding terrain failed to detect a systematic difference. This indicates that similar environmental modification in the vicinity of rural reporting stations should not greatly influence the observed temperature.

The frequencies of subzero Fahrenheit (<–17.8°C) temperatures at several stations were compared. The most frequent occurrences were in small basins, and

the least frequent on ridges. The frequency of occurrence of subzero temperatures along the Connecticut River increased with increasing snow cover. Most occurrences of morning temperatures less than –20°C coincided with snow cover exceeding 20 cm.

We propose that the mean annual occurrence of temperatures less than 0°F (–17.8°C), which is cataloged in NOAA (National Oceanic and Atmospheric Administration) climatologies, is a reliable index of frequency of strong winter inversions over snow-covered terrain. The geometric standard deviation of temperature difference in a locality provides an index useful in locating persistent inversion elevation.

Local Variation in Winter Morning Air Temperature

AUSTIN W. HOGAN AND MICHAEL G. FERRICK

PART I: A RIVER SURFACE REFERENCE

INTRODUCTION

Many problems in engineering, hydrology, and operational meteorology are further complicated by local variation in winter surface air temperature. There are additional problems in preparing dispersion estimates over snow-covered terrain, which are related to the alteration of vertical temperature structure that accompanies these surface temperature variations. Locally severe ice damage may occur when subfreezing air remains in valleys as warm advection produces rain above. Lakes and waterways may suffer early freezeup in the windless conditions that characterize strong local inversions. More complex problems arise when estimates of frost depth, extreme temperatures, or the frequency of strong local inversions are required at a proposed operational site when only data from dissimilar reporting stations are available.

An example of the magnitude of winter temperature variation in the 42°N to 45°N latitude segment of the eastern United States and adjacent Canada is shown in Figure 1. This figure illustrates the mean annual number of subzero days, having temperatures less than –17.8°C (0°F), from NOAA (1982) *Climatography* no. 20, *The Climate of Cities,* for the period 1951–1980. The Canadian data was calculated from cooperator station records for the same period by Dr. David Phillips.* Note the complexity of the frequency isopleths in northern New York, Vermont, and New Hampshire in Figure 1. The frequency of subzero (–17.8°C) temperature occurrences varies by factors of two over distances of less than 30 km. The complexity and scale of the subzero frequency isopleths are comparable to those of cold day isotherms of the island of Hokkaido, given by Nakamura and Magono (1982). The variation in frequency of subzero occurrences in the vicinity of Stillwater Reservoir, New York (designated by the plotted 45), is well known in that area, according to L. Lansing.† The complexity and number of isopleths in Figure 1 are probably diminished by the sparsity of reporting stations. This may induce unexpected problems if an activity dependent on operational, engineering, or air pollution meteorology is undertaken in this area.

The area in Figure 1 bounded by 42°–45°N, 70°–80°W, probably experiences a similar quasi-uniform distribution of tropospheric air masses over a 30-year period. An inspection of the station tabulation of NOAA (1982) is sufficient to conclude that simple elevation difference cannot account for these great differences in subzero (<–17.8°C) temperature frequency. A comparison of 0700 (1200 UTC) winter temperatures observed at Mt. Washington, New Hampshire,** at elevation 1,920 m to those observed in the Connecticut River Valley at Z elevation 230 m, less than 60 km distant is shown in Figure 2. The locations of Mt. Washington and Z are noted on Figure 1. The lower elevation temperature is less than the potential temperature and often absolutely lesser than the mountain station. This systematic difference can be partially accounted for by the decoupling of the surface layer (Miller 1956a, b) from the troposphere by one or more inversions

* Personal communication, Dr. David Phillips, Atmospheric Environment Service (AES), Environment Canada, 1993.

† Personal communication, L. Lansing, Cooperative Observer, National Weather Service, Boonville, New York, ca. 1970.

** Personal communication, K. Rancourt, Mount Washington Observatory, 1991–92.

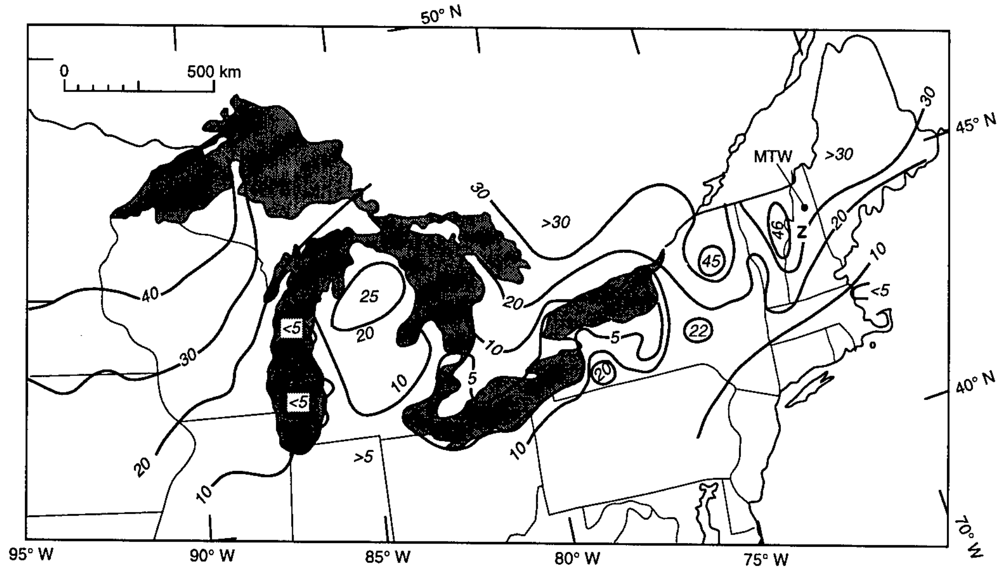


Figure 1. Mean annual number of subzero (<−17.8°C) days occurring during the period 1951–80, from NOAA (1982). The Canadian data were provided by D. Phillips, AES, Downsview, Ontario. The study area of Figure 3 lies along the Connecticut River near Z.

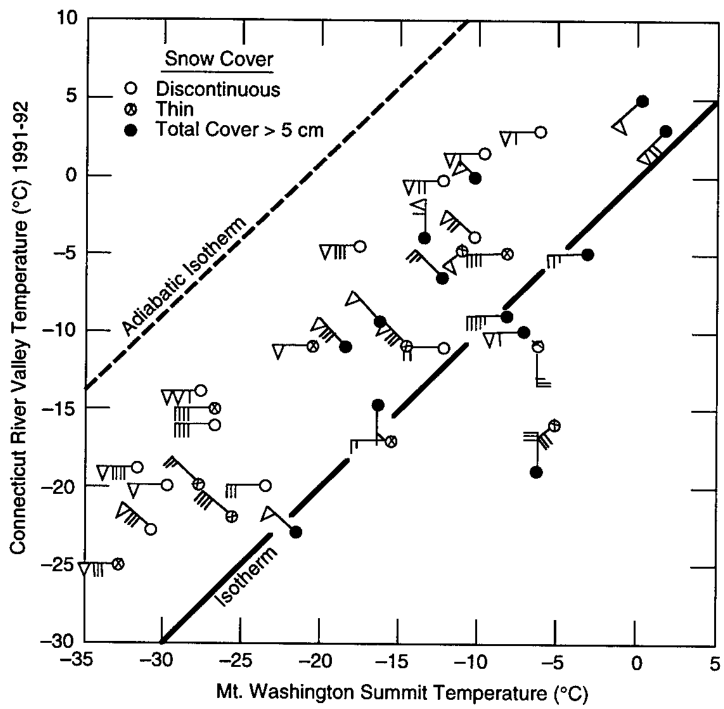


Figure 2. Comparison of 07LCL(12UT) temperatures observed at Mt. Washington and Piermont, New Hampshire, during the winter 1991–92. The station circles denote snow cover at Z; the wind barbs denote wind speed (5 m/s per barb) and direction at Mt. Washington.

produced by the radiational heat loss at the snow/air interface. The magnitude of this decoupling effect may be influenced by local features, causing some of the observed variation in local temperature.

There are many examples in literature of diminution of surface air temperature over snow cover reviewed in Munn (1966). Large local temperature variations have been observed in Alaska by Benson and Bowling (1973) and in Hokkaido by Magono et al. (1982). Namias (1985) suggested synoptic scale influence of snow cover on temperature and precipitation. Dewey (1977) showed that the presence of snow cover locally influenced daily maximum and minimum temperature. Longley (1949) analyzed a 100-year climatic record at Quebec, and showed the spring increase of morning minimum air temperature to stagnate at 0°C for several days, until the snow cover diminished. The addition of the influence of snow-

covered ground to the terrain related microclimate concepts of Geiger (1965) poses several questions relative to a priori estimates of the temperature structure of overlying air.

A RIVER PLANE TEMPERATURE REFERENCE

We propose that the air over a still river provides a relatively uniform surface reference plane that can be used to compare air temperatures occurring over differing surfaces near the river on most winter days.

The Connecticut River is impounded at Wilder Dam. The impoundment is at 117 m (384 ft) above mean sea level and this level remains constant along the watercourse of Figure 3. The water in this impoundment is replaced over a three-day period by bottom water from McIndoe Falls Dam approximately 30 km upstream of Figure 3. The river is from 100 to 200 m wide along this section of its course; the lower basin (terrain less than 25 m above water level) along the river varies in width.

The frequent recharge of the river prevents stratification or stagnation of its water, which might cause a variation in heat source strength among stagnated pools. The river is frozen and snow covered on most winter days. This combination of spatially uniform subsurface temperature with a level snow surface free of obstacles provides an approximation of a "reference" radiating snow surface with a relatively constant roughness length (Stull 1988) that is 50 km in north-south extent. Perovich* has recently observed that a few millimeters of snow on ice presents a uniform radiative plane representative of a snow surface.

* Personal communication, D. Perovich, CRREL, 1994

The reference plane is a layer of air just above the lowest elevation in the observation area, and it can radiate to a larger sector of sky than a point along a slope or in a smaller basin. Cold air does not drain from the reference plane, but the air temperature can continue to diminish through exchange with the surface and radiation to the sky. Comparison of nearby surface temperatures in other terrain settings with those of the river plane can initiate understanding of physical processes responsible for surface temperature variation. On a few days each winter, the river may present a surface of open water, an alternation of ice and water, or a thickening snow-free ice plane. These conditions provide an opportunity to study some additional mechanisms responsible for local air temperature modification, and will be discussed in subsequent sections.

The winter surface is variable with properties greatly dependent on the thickness and continuity of snow cover. Quantitative snow measurements were made at Z, and qualitative snow observations were made at several other points each day. Snow depths were estimated with respect to fixed objects as done in Hokkaido, and surface cover was visually estimated from the relative area of bare spots. We propose, after comparison of the quantitative and qualitative observations, that the observation at Z numerically characterizes the snow cover along the Connecticut River Valley during the core of winter. This characterization is shown in Table 1. We will stratify data, using the categories in Table 1, to examine processes that produce variation in winter morning air temperature in the study area.

Table 1. Definition of snow depth categories.

< 1 cm	This is reported for some early winter observations and some days following thaws when snow is predominantly absent from the basin. It is also the characteristic of a light dusting of bare ground by flurries.
< 3 cm	This is associated with sparse, or discontinuous remnant, snow cover in the Connecticut River Valley. The <1 cm category is included within this category for analysis.
< 10 cm	This is associated with general snow cover, but bare or sparse spots occur in fields and on southern exposures. Grasses and field crop stubble may protrude through 10 cm of snow.
< 20 cm	This category was extracted from the data set to test the statistical robustness of some analyses.
< 30 cm	This amount is associated with nearly total and continuous snow cover.
> 30 cm	This amount is associated with total and continuous snow cover at all locations.

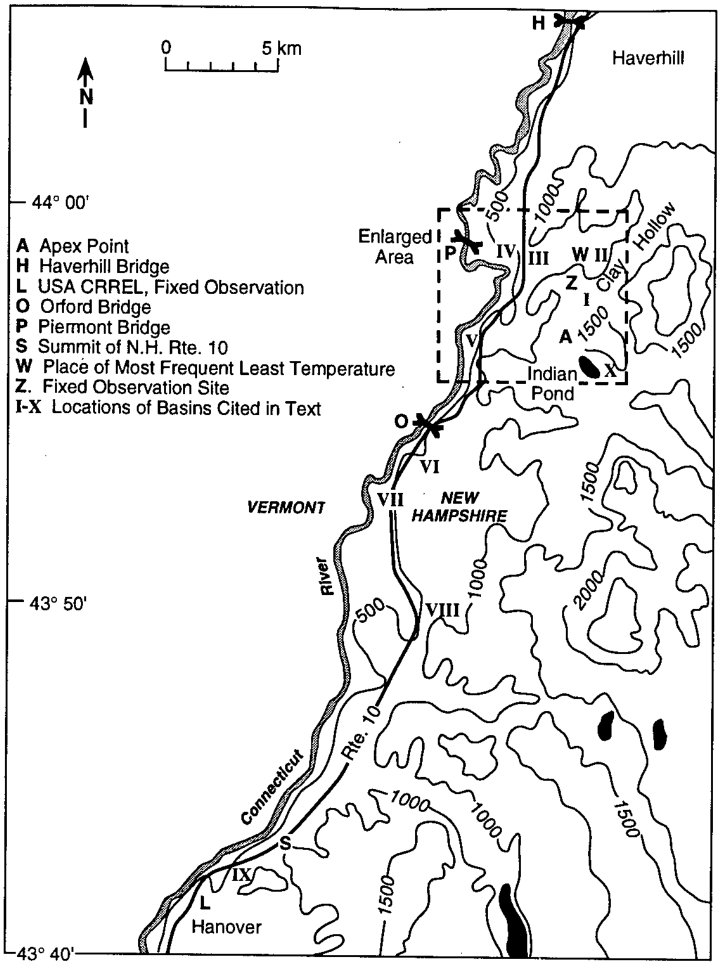


Figure 3. Topography of the study area east of the Connecticut River (left panel). The locations of observation stations are designated on the right panel. The alphanumeric point designators, roman numeral basin designators, and place names coordinate all figures and tables in Parts I, II, and III. "Enlarged area" is shown in detail in Figure 9.

Baker et al. (1991, 1992) investigated the depth of snow necessary to influence albedo over agricultural land; 5 cm snow was sufficient to mask bare ground, but 15 cm was required to mask alfalfa. Baker et al. (1992) noted a systematic reduction of air temperature over snow cover with respect to bare ground, and that 10 cm of snow cover was accompanied by 8.4°C diminution in mean temperature. Kung et al. (1964) considered 5 in. (12.5 cm) of snow to be sufficient to achieve a uniform snow albedo. These works were considered, in conjunction with local experience, in the formulation of Table 1. The snow depth and cover parameters defined in Table 1 will be used with air temperature parameters to objectively examine air temperature-snow cover interaction. The <1-cm and < 3-cm snow depth categories are combined for the analyses, as bare ground is dominant in both cases. The 3- to 10-cm and 10- to 30-cm categories approximate the Baker et al. and Kung albedo criteria. Additional 10- to 20- and 20- to 30-cm depth categories are used in the analyses to examine the consistency of the data and the applicability of albedo generalization in this terrain. Most of the days with >30 cm of snow occurred in February during the 1990-1993 winters. This is considered in comparison with the Magono et al. (1982) and Maki et al. (1986) Hokkaido experiments above snow cover which usually exceeded 1 m in depth.

The location of the experiment is noted by Z in Figure 1. Observation and measuring sites used in this research are plotted on a map in Figure 3. Points cited in the text are named or alphanumerically designated for coordination with other figures and maps in Parts II and III. Fixed recording sites, measuring sites, and important features are designated by capital letters. Small basins surrounding observation points are designated by Roman numerals. Three hamlets, Piermont, Orford, and Lyme, are noted by name. These will be considered in defining the reference temperature along the river plane.

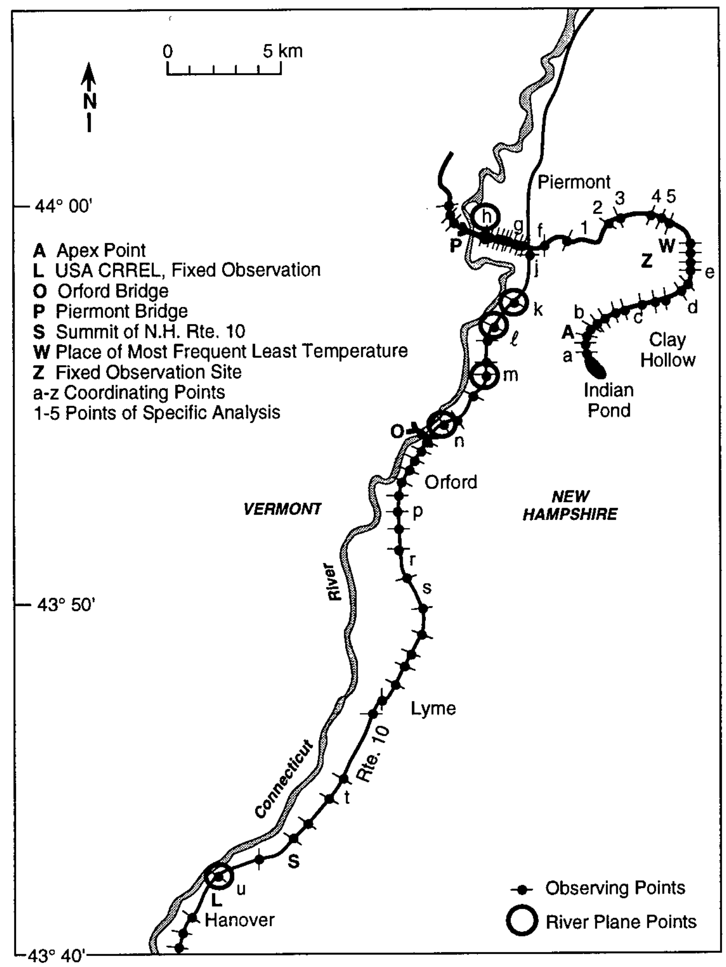


Figure 3 (cont'd).

INSTRUMENTATION, OBSERVATIONS, AND MEASUREMENT TECHNIQUES

It is extremely difficult to precisely measure ambient air temperature in near calm conditions over a snow surface as noted by Andreas (1986). Aspiration of the sensor may disturb the stability of the layer being measured, and the presence of an observer or signal processing equipment may disturb the heat budget of the sensitive volume. The sensor must be radiationally shielded from both sky and snow, and because of the large temperature gradient above snow-covered ground, the distance between the sensor and the snow surface must be precisely maintained, without disturbing the surface in the vicinity of the sensor.

Maintenance of calibration, sensor height, and surface integrity of 100 temperature sensors in the area shown in Figure 3 would be a massive undertaking. Additionally, access to all points of interest might not be possible as the multiple owners may not all be amenable to hosting observation stations. We propose that an aspirated moving probe is capable of making continuous near-surface temperature measurements, which can be scaled to yield horizontal and vertical temperature profiles.

A diagram illustrating the technique for analyzing horizontal and vertical temperature gradients is shown on a cross section of a hypothetical small basin in Figure 4. The smoke from the house chimney in the diagram flattens as an indication of temperature inversion at that elevation. A series of temperature measurements along a transect of the basin *A-B-C* would produce the spatial variation in temperature shown below the cross section. Plotting the temperature as a function of measurement elevation would yield the vertical profiles *AB* and *CB* shown to the left of the diagram.

All considerations of the data, and subsequent analyses of the data sets or subsets, would have been first stratified by determining if the near surface air temperatures in the study area ap-

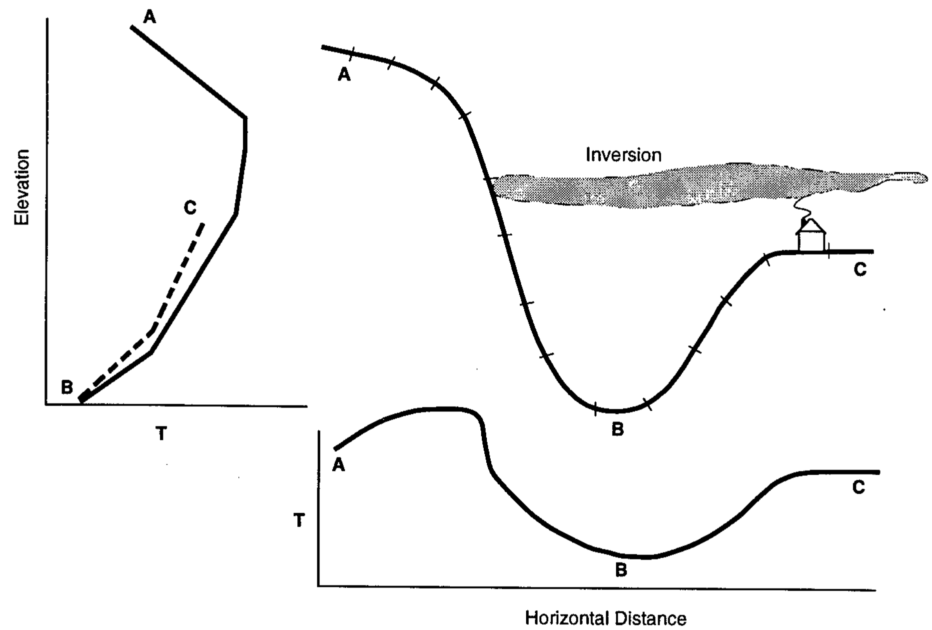


Figure 4. Diagram indicating method of transforming surface temperature observations to vertical temperature structure.

proximated lapse or inversion structure at the time of observation. This was determined by comparing the temperature at W in Figure 3 about 100 m above river level, and the temperature at A, 300 m above river level, the apex of observation. A day that has a temperature at A exceeding the temperature at W by any amount is an inversion day. A lapse day (not necessarily adiabatic) occurs when the temperature decreases above the elevation of W. Individual layers with lapse or inversion structure may be present in either case, but this analysis stratifies the data with respect to the observed temperature profile and does not use a temperature in our reference plane as a criterion.

A shielded, relatively rapid response thermistor (YSI 400 with specified $0.63\Delta T$ response of 10 s to a step change in air temperature) and its digital readout (0.1°C) were adapted with convenient mountings for a car or truck. The melting point temperature was verified in the field by wetting ice on the surface of the thermistor. Some experimentation was necessary to select mountings 1.25 m above the ground level that were not influenced by radiation or advection of heat from the vehicle; the mirror mount finally selected was found to reflect the free stream temperature whenever the vehicle was in motion. The response parameters are tabulated in Table 2, and define the time and space resolution of the temperature measurements reported in this work.

These are derived from the manufacturer's 10-s time constant and a minimum aspiration speed of 5 m/s to eliminate self heating and the possibility of convective heating by the vehicle. This was experimentally tested by comparing the observed temperature achieved after a 50-m roll at 5 m/s with the nearby free-air temperature measured by slinging an identical thermistor. The tabulation indicates that it is possible to examine temperature changes on a less than 100-m spatial scale on the sparsely traveled rural roads, which constitute most of the experimental area, and with 200- to 300-m space resolution elsewhere.

A question may arise regarding modification of the temperature field by other passing vehicles. This was not apparent during measurements, but an experiment was conducted to estimate the magnitude of temperature modification that may be induced. On several still days with surface temperature of less than –20°C, the probing vehicle

Table 2. Distance required to respond to a change in temperature.

Vehicle speed (m/s)	*Distance (m) to recover 0.95(ΔT)*
5	60 (minimum aspiration rate)
8	100
16	200
25	310

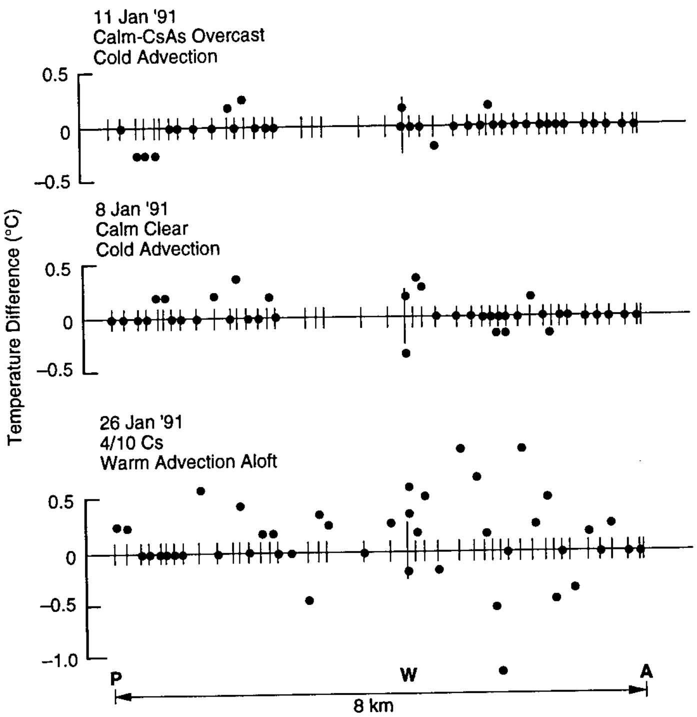


Figure 5. Magnitude of difference of repetitive air temperature measurements (solid circles), accompanying a direction reversal, over a 3- to 12-minute period. Observed differences of less than 0.1°C are plotted along the zero axes.

was brought to a halt on a deserted road while noting the air temperature. The vehicle was then promptly reversed through its exhaust plume, and the temperature again noted. Temperature increases of 0.5 to 1.5°C could be achieved in this way. This was easily avoided by maintaining a distance of 100 m from the infrequent vehicles encountered along the path of observations. Any temperature increase due to this modification necessitated by closer proximity to vehicles while passing through populated areas is considered as part of the "heat island" modification.

Additional experiments were conducted to examine the "sampling error" or "position error" of the observations. These experimental observations were conducted by first making observations at several points along the 8-km transect noted by *P–W–A* in Figure 3, and then repeating the observations in reverse order. This reversal altered the path followed by the probe and its proximity to banks and barriers along the course. It also reversed the sign of the rate of temperature change when penetrating inversions. The elapsed time between repetitive measurements at individual points was a minimum of 3 and a maximum of 12 minutes. The results obtained on three days with differing meteorological conditions, representing those most frequent in the area, are given in Figure 5.

Lapse (but not necessarily adiabatic) temperature structure is sometimes present in the Connecticut River Valley when cold advection occurs in winter. The upper two temperature difference plots in Figure 5 show the difference between replicate temperature measurements at individual points, as described above, observed during cold advection with apparent lapse structure along the slope. Inversion is frequently present in this valley when warmer air is advected over it in winter. The lower plot in Figure 5 shows the temperature differences from replicate measurements under warm advection and apparent inversion conditions. Lapse and inversion were defined by *W–A* temperature difference.

Precise (within the 0.1°C resolution of the digital display) replication of temperature observation was achieved in 56 of 77 instances under lapse and cold advection. The maximum difference observed was 0.3°C, and four observing points showed temperature differences greater than 0.1°C on both occasions. This indicates that this technique of temperature measurement does not induce apparent temperature variation of greater than 0.1°C. However, position, sampling, and temporal temperature differences of as much as 0.3°C may occur in the observed data.

Precise (0.1°C) replication of temperature observed was only achieved in 16 of 44 instances under inversion conditions. Differences exceeding 1.0°C were observed, and differences of 0.3 to 0.5°C were common between replicate observations. These differences are attributable to the time response of the temperature sensor, and the reversal of the sign of the rate of change of temperature when penetrating inversions. Subsequent sections will show these differences in measurement precision are tolerable when compared to the temperature differences in the vicinity of these inversions. More realistically, the sensor probe advances at about 12 m/s along a 10% (100 m/km) grade. The distance response of Table 2 shows the sensor responds to 0.95 of an temperature change in 150 m, which is 15 m of elevation change along the slope. A one-way penetration can then detect inversions of 15-m thickness; a replicate penetration can determine the temperature at this inversion within 1°C.

The observations obtained for the analysis of local temperature differences were obtained in conjunction with other research being conducted in this area but no meteorological preselection occurred. The observer was not present in the study area on many days each winter. The statistical analyses that follow are based on 158 December, January, and February days during the period December 1990 through February 1993. All observations were made prior to local sunrise. It was not always possible to complete both an east–west and north–south transect on some days prior to sunrise, limiting the data set to 100–120 days for some comparisons. No days have been deleted from the data set. Rain or freezing rain truncated data collection on some days, and one observation attempt is incomplete due to inaccessibility of point A due to blowing snow. Data from other periods during the winters 1988–1994, obtained with the same techniques, are used to illustrate some specific cases.

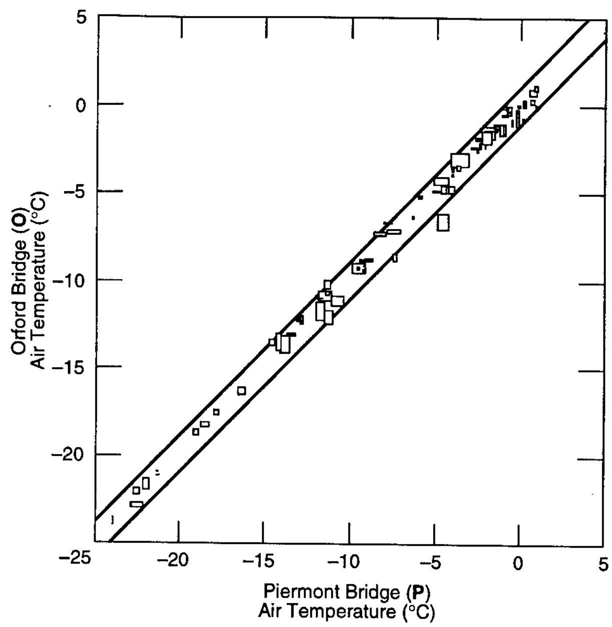


Figure 6. Temperature differences measured on two bridges 9 km apart along the Connecticut River. The plotted boxes enclose the range of temperatures observed on each bridge.

EVALUATION OF THE CONNECTICUT RIVER SURFACE AS A TEMPERATURE REFERENCE PLANE

We propose that the air over a still river can be used as a reference plane for comparison of other air temperatures. The air temperatures above the river, and along a plane coincident with the river bank, need to be examined to determine the contiguity of this surface as a temperature reference.

A preliminary examination of the variation in temperature on two bridges, located 9 km apart at P and O in Figure 3, was made by Hogan and Ferrick (1991). Figure 6, adapted from that work, compares the range of temperatures observed each morning on the Orford–Fairlee (O) bridge to the range of temperatures measured on the Piermont–Bradford (P) bridge a few minutes earlier each day. The temperature difference exceeded 1°C on only 3 of 62 days of observations. On most days, the variation along the length of a bridge, or the variation among replicate observations, equaled or exceeded the temperature difference between the two bridges.

An additional bridge lies 10 km north of the main experiment area, at H in Figure 3. A brief series of experiments were conducted in January and February 1992 to examine the temperature differences along this 20 km of the river plane. Observations were completed on 15 mornings and the relative frequency of lapse and inversion conditions approximated that of the entire (1990–93) experiment period. The data are tabulated in Table 3, which indicates that an absolute difference of (1°C) in air temperature was exceeded only once

Table 3. Temperature differences among three bridges.

Snow depth (cm) (temperature structure)	*Temperature difference (°C) (absolute value)* *O, P*	*P, H*	*O, H*
<3 (Lapse)	0.2	0.0	0.2
<10 (Lapse)	0.1	0.1	0.2
	0.5	1.3	0.8
<30 (Lapse)	0.1	0.5	0.4
	0.7	0.3	1.0
	0.4	0.4	0.0
<3 (Inversion)	0.5	0.0	0.5
	2.3	0.3	2.6
	0.5	0.0	0.5
<10 (Inversion)	1.8	0.6	1.2
	0.3	0.1	0.2
<30 (Inversion)	0.8	1.0	0.2
	0.0	1.2	1.2
	0.5	0.3	0.2
	0.8	1.0	0.2

on days with lapse temperature structure. Air temperature differences among the bridges exceeded 1.0°C on 3 of 9 days with inversion structure. It is interesting to note that the temperature differences do not seem systematic, and that the maximum temperature difference occurs between the two most distant bridges on only two occasions.

We can measure the temperature over the river at an elevation of a few meters above the river only on those bridges. At seven other points (noted on Fig. 3), air temperatures can be measured at the same elevation along the bank, 10 to 100 m from the river's edge. The northernmost of these points (*h*) is 100 m east of bridge *P*. A comparison of the extreme differences among six independent temperature observations on the bridge *P* and two replicate temperature observations on the bank 100 m east of the bridge portal has been made to investigate the magnitude of difference among these observations.

Table 4. Frequency of temperature difference on (*P*) Piermont bridge.

Stability	*Difference among 141 sets of six observations (°C)*		
	Median	84th Percentile	Maximum
Lapse	<0.1	<0.1	0.4
Inversion	<0.1	<0.2	1.4

Air temperature was first measured on the east bank, then a few seconds later at the same height at the east portal, center, and west portal of the bridge; about 200 s later the measurements were repeated in reverse order. The mean temperature and the maximum temperature difference were calculated for each set of six observations along the bridge. The temperature differences have been separated according to presence of lapse or inversion conditions and stratified into the snow depth categories defined in Table 1. An analysis of the frequency of greatest temperature difference observed on the bridge is tabulated in Table 4. More than two-thirds of all observations have extreme differences of less than the one digit uncertainty of the readout device used to measure the air temperature. This poses some problems in the frequency analysis. Three readable temperature difference intervals (±0.0°, ±0.1° ±0.2°C) lay within the digital repeatability of comparison of 0.2°C. We have used the observed temperature difference in these analyses, rather than grouping all differences of less than 0.2°C in a single interval.

The mean of these six measured temperatures is compared to the mean of the two temperatures measured 100 m distant in Table 5, again separated according to lapse or inversion and stratified by snow category. The difference in mean temperatures rarely exceeds ±0.1°C in lapse con-

Table 5. Frequency of temperature difference at a 100-m distance from bridge.

Snow cover and stability	*Relative frequency of occurrence in temperature difference class* 0°C	<0.1°C	<0.2°C	<0.4°C	<0.8°C	<1.6°C	*Number*
<3 cm, lapse	0.73	0.27	0.0	0.0	0.0	0.0	15
3–10 cm	0.63	0.31	0.06	0.0	0.0	0.0	16
10–30 cm	0.36	0.50	0.07	0.07	0.0	0.0	14
>30 cm	0.67	0.0	0.33	0.0	0.0	0.0	3
Cumulative	0.57	0.91	0.98	1.00			48
<3 cm, inversion	0.41	0.27	0.09	0.05	0.14	0.05	22
3–10 cm	0.29	0.50	0.13	0.04	0.04	0.0	24
10–30 cm	0.31	0.23	0.23	0.10	0.07	0.05	39
>30 cm	0.13	0.00	0.13	0.13	0.38	0.25	08
Cumulative	0.31	0.71	0.76	0.84	0.95	1.00	93
Cumulative all	0.40	0.49	0.86	0.88	0.96	1.00	141

ditions, and fewer than one-third of all differences exceed this magnitude. About two-thirds of differences exceeding ±0.1°C were observed when inversion was present over more than 10 cm of snow cover. This analysis indicates that air temperatures measured within 100 m of the river may be indicative of the air temperature above the river surface, but that differences may increase under inversion when the surface is uniformly covered with snow. Two measurements of air temperature within 100 m of the river numerically approximated six air temperature measurements above the surface of the river. The differences are within the resolution of the digital readout.

The difference between the surface temperature observed at the seven near river points and the mean temperature measured along bridge *P* has been used to construct a frequency analysis of temperature difference as a function of distance along the river plane. A frequency analysis of the 141 morning air temperatures observed on bridge *P* showed them to approximate a normal distribution in the interval 2 > T > –26°C.

The temperature differences with respect to *P* observed at the six points within 10 km of *P* were stratified with respect to stability and pooled in two (lapse and inversion) data sets. When the logarithms of a variable are normally distributed with respect to number of occurrences, the ratio of the variable at the 84th and 50th percentile of occurrence is equal to the geometric standard deviation (GSD) of the variable. This method is well described in Hinds (1982) and permits graphical extraction, or approximation, of the median and geometric standard deviation from relatively sparse data. About 40 observations under lapse, and 80 under inversion, are available for comparison of temperatures along the river plane to the bridge temperature. Less than the total 158 data days are used in this analysis, as temperature measurements along the north–south extent of the experiment area were sometimes not completed before sunrise.

Preliminary plots of the logarithm of the absolute value of temperature difference versus number of occurrences showed that the logarithms of the temperature differences were normally distributed under lapse conditions. The logarithms of more than 68% of the temperature differences were normally distributed under inversion conditions.

The median and 84th percentile of temperature difference have been extracted for lapse and inversion conditions at the individual observing points along the river, and are listed in Table 6. The lower case letter designators coordinate these points with other places to be cited in Parts II and III. The data for the point 100 m from the bridge are repeated for reference in this table, but the standard deviation is not calculated for this point in the following figures, as its 84th and 50th percentiles lay within the digital resolution of the display.

Table 6. Frequency analysis of temperature difference along the river plane referenced to the temperature at Piermont bridge.

		Temperature difference (°C)			
		Under lapse		*Under inversion*	
Symbol (Fig. 3)	*Distance (km)*	*median*	*84%*	*median*	*84%*
h	0.1	<0.1	<0.1	<0.1	0.2
k	2.6	<0.1	0.3	0.3	0.8
l	5.0	0.2	0.4	0.5	1.0
m	5.9	0.2	0.5	0.6	1.0
n	7.7	0.3	0.6	0.5	1.1
O	8.2	0.2	0.6	0.5	1.2
u	29.4	0.6	1.2	1.0	1.6

The median and geometric standard deviation of the temperature differences at these seven near-river points, with respect to the Piermont bridge (*P*) temperature, are plotted as a function of distance south of *P* in Figure 7. The coordinator letter is noted on the distance scale. The 84th percentile of temperature difference, and the cumulative frequency of lesser temperature at each point, are similarly plotted in Figure 8.

From these analyses, air temperatures above the relatively uniform surface of the frequently recharged Connecticut River impoundment shown in Figure 3 were found to differ by less than 1°C on most winter mornings. Critical analyses of Figures 7, 8 and Tables 4, 5, and 6 indicate temperature difference systematically increases with distance from the reference bridge. Temperatures within 10 km of *P* are quasi-uniformly distributed between colder and warmer with respect to that at *P*; those at 30 km are warmer than at *P* most of the time. These analyses indicate that the plane intersecting the bank of the Connecticut River provides a relatively uniform reference plane, with respect to winter morning air temperature, along the segment of the Connecticut River Valley shown in Figure 3.

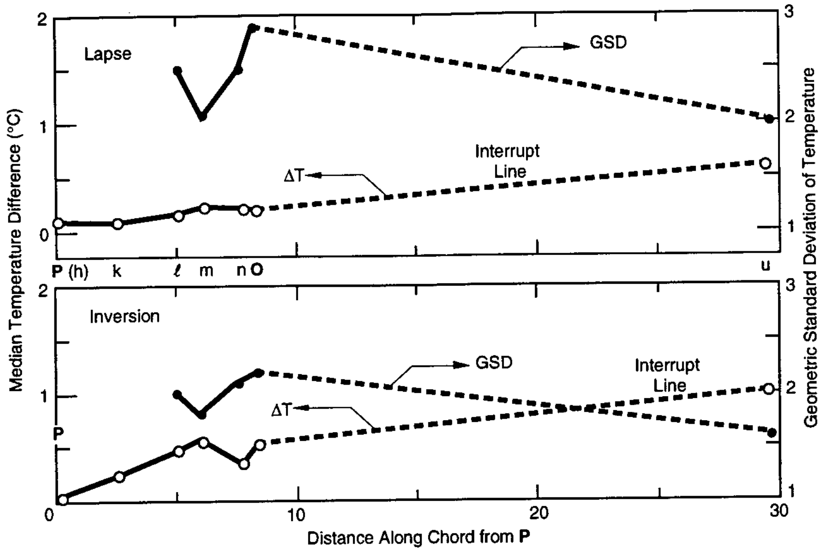


Figure 7. Median temperature difference and the geometric standard deviation of temperature difference measured along the river under lapse and inversion conditions. The letters coordinate the location of observing points with Figure 3.

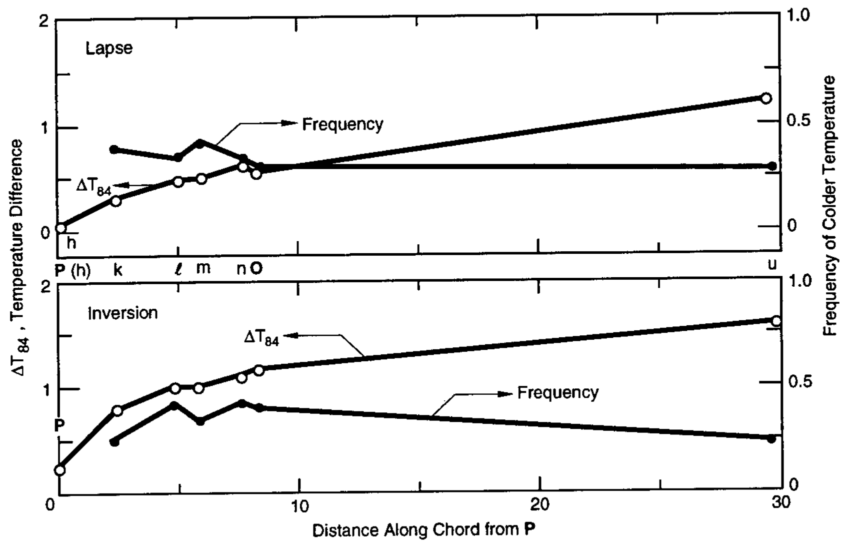


Figure 8. Eighty-fourth percentile of temperature difference, and the frequency of colder air temperature than at P, the reference point.

The data were initially stratified by both snow depth category and temperature structure to perform the analyses shown in Tables 4, 5, and 6. This stratification was deleted to simplify the presentations in those tables. The frequency of temperature difference greater than 0.4°C at three points along the river at less than and greater than the 10 cm snow cover which Baker et al. (1992) considered necessary to mask underlying surfaces are listed in Table 7. Lapse and inversion data are combined in Table 7, and the temperature difference at the point 30 km south of bridge P, exceeds 0.4°C three-fourths of the time, with all snow conditions. There is a systematic increase in the frequency of greater temperature difference when more than 10 cm of snow cover is present at the points along the river.

Table 7. Trend of temperature difference relative to snow cover.

Symbol	Distance (km)	Frequency of >0.4°C temperature difference <10 cm snow	>10 cm snow	>30 cm (3 cases)
h	0.1	0.06	0.13	All
k	2.6	0.22	0.31	All
l	5.9	0.40	0.61	All

DISCUSSION

The NOAA and AES climatic records show that relatively great differences in the frequency of subzero (–17.8°C) air temperatures occur over small horizontal distances in many places in the

northeastern United States and eastern provinces of Canada. We found in preliminary analysis that lesser temperature differences occurred on winter mornings at river level than on a nearby mountain, and we seek the physical mechanisms responsible for this. A systematic approach to temperature observation was developed, allowing the temperature to be repetitively observed adjacent to landmarks of known elevation. We have proposed that the vertical temperature structure of the air within the valley can be extracted from temperature observations at two elevations. We cannot independently verify this with free air observation, but the inferred lapse and inversion structure is consistent with other observations. We have shown that an early morning air temperature survey can be quite quickly accomplished in a 10- × 30-km area of 330-m local relief. Analysis of the observed temperatures in the plane coincident with the bank of the Connecticut River provide a temperature reference plane, which differs by less than 1.2°C over distances of 8 km (and less than 1.6°C over distances of 29 km) on most winter mornings. This measured temperature uniformity indicates that the radiative properties of the surface along the river and the atmosphere above it are relatively uniform over this distance.

Magono and his coworkers used a large array of maximum-minimum thermometers and several tethersonde surveys to examine the temperature distribution and vertical temperature structure in the Moshiri Basin. We have used a moving probe, as Benson and Bowling (1973) did, to survey temperature difference over a larger area. Our results are essentially in agreement. This indicates that the moving probe technique of examining temperature structure over hilly terrain, proposed in Figure 4, was apparently capable of producing the same results as a large field of thermometers and a sonde. We used this apparent lapse or inversion structure, which was independent of observations in the river basin, to stratify our analysis of temperature variation along the river.

Several techniques of frequency analysis were used in determining the dimensions of a region of relatively uniform air temperature above the Connecticut River. These techniques were applicable to the analysis of this data field, because we found the logarithms of temperature differences along the river to be normally distributed under lapse, and nearly so under inversion. Hatch and Choate (1929) showed that the slope of the central portion of the log normal distribution (i.e., the ratio of the variable at the 84th percentile to that at the median) was numerically equal to the GSD of the variable. We propose that the GSD of temperature difference is a sensitive indicator of local differences in air exchange, where surface air temperatures are lower than the isentropic extension of tropospheric temperatures on a majority of winter mornings. The frequency analysis of temperature differences may provide an economical approach to determining exchange and dispersion characteristics in an area of sparse meteorological record, where activities or operations may require meteorological support. This proposal will be tested in Parts II and III.

CONCLUSIONS OF PART I

The logarithms of the temperature differences observed with respect to a reference point temperature were normally distributed along 30 km of the Connecticut River Valley. The median and geometric standard deviation of the logarithms of temperature differences can be calculated and considered in evaluating the river plane as a reference temperature field. The median value of temperature difference increased in an orderly way as distance from the reference point increased. This difference increased in magnitude as snow cover increased in depth and continuity.

We conclude that atmospheric influences on outgoing radiation were relatively uniform over this 30-km segment of the Connecticut River during this experiment period, allowing air temperatures to achieve this uniformity. This analysis has defined the resolution of a moving probe air temperature measuring system and the homogeneity of surface air temperatures along the Connecticut River plane sufficiently to allow the river plane temperature to be used as a reference for analysis of temperature-modifying processes in differing terrain along the river.

PART II: OTHER TEMPERATURE VARIATIONS WITH RESPECT TO THOSE ALONG A RIVER

INTRODUCTION

Part I introduced the concept of the using the air temperature observed along the plane coincident with the bank of the Connecticut River, near 44°N latitude, as a reference temperature to examine the processes responsible for large morning air temperature differences observed in the region. The logarithms of temperature difference among points along this plane were normally distributed, and the geometric standard deviation of temperature difference was quite similar along a 30-km segment of the river valley. Temperature differences of the greatest magnitude occurred most frequently under inversion and over deepest snow cover. This report will apply the techniques of measurement and analysis described in Part I to examining the much larger temperature differences observed in hilly terrain east of the Connecticut River.

It is well known, and summarized by Clements (1989), that basins and valleys experience overnight air temperatures that are lower than those found at nearby ridges or extensive plains. Miller (1956a,b) studied the influence of snow cover, intercepted snow, and vegetation on air temperature in the Sierra and Rocky Mountains, showing a decoupling of the surface layer. We propose that principles of polar and mountain meteorology can be synthesized with smaller distance scale concepts of nocturnal inversion formation by Andre and Mahrt (1982), Arya (1981), Yamada (1979), and Sutherland (1980), and the winter basin flow concepts of Maki and Harimaya (1988), to examine the differences of surface air temperature observed in the Connecticut Valley by Hogan and Ferrick (1990).

Nakamura and Magono (1982) showed that air temperature just above a snow surface diminishes with wind speed, but that the air temperature at conventional screen height of 1.25 m becomes a minimum when wind speed 9 m above the surface is about 1 m/s. This dependence of the "surface" (i.e., 1.25-m) air temperature on wind speed a few meters above apparently contributes to the production of "surface" temperature differences of several degrees over horizontal distances of a few hundreds of meters in snow-covered terrain. It also induces "surface" temperature variation in conjunction with slope and surface roughness, which modifies lapse temperature structure along or in the vicinity of hills or mountains. Although Geiger's (1965) descriptions of drainage winds, sink holes, and cold valleys are well respected, there are significant differences of opinion regarding the relative contribution of cold air drainage and resident stagnation to diminished air temperatures in basins.

Magono et al. (1982) quite successfully modeled the horizontal distribution of winter temperatures in Hokkaido, but were not able to model the extremely cold air found in small basins and narrow valleys on the coldest days. They conducted an additional field experiment in the Moshiri Basin, which they proposed produced the coldest air in Hokkaido. They found an inversion to form more rapidly in the basin, which decoupled the basin from the down valley winds that continued to circulate heat through the night along the adjacent slopes. Maki et al. (1986), Kondo (1986), and Maki and Harimaya (1988) studied and modeled the heat budget in this and other complex Hokkaido basins. Their models predict nocturnal winter cooling to be greater in basins than on flat terrain, and also that nocturnal winter cooling will be greater at the base than at the tops of mountains. The mountain (or hill) tops are not as frequently decoupled from tropospheric heat advection. Downslope drainage may also induce circulation of this warmer air along slopes.

Geiger (1965) provided marvelous insight to local climate variation. The observation, measuring, and experiment sites designated in Figure 9 reflect these, and later, ideas of Geiger (1965, 1969). The measuring points were chosen to include several apparent cold air drains and locations similar to those representative of microclimates given by Geiger, but some more specific recent studies dominated the experiment plan. Measuring points reflecting hamlets (Landsberg 1981), basins (Maki et al. 1986), slopes (Maki and Harimaya 1988), barriers (Baines 1979), and forest/field boundaries (Raynor 1971) were selected to examine the influence of these cultural, topographic, and natural surface features. Raynor's (1971) boundaries and Baines' (1979) barriers to flow in the form of narrowing or widening of tributary valleys, small ridges, woods, and forests were used to define the boundaries of some slopes and basins for analysis.

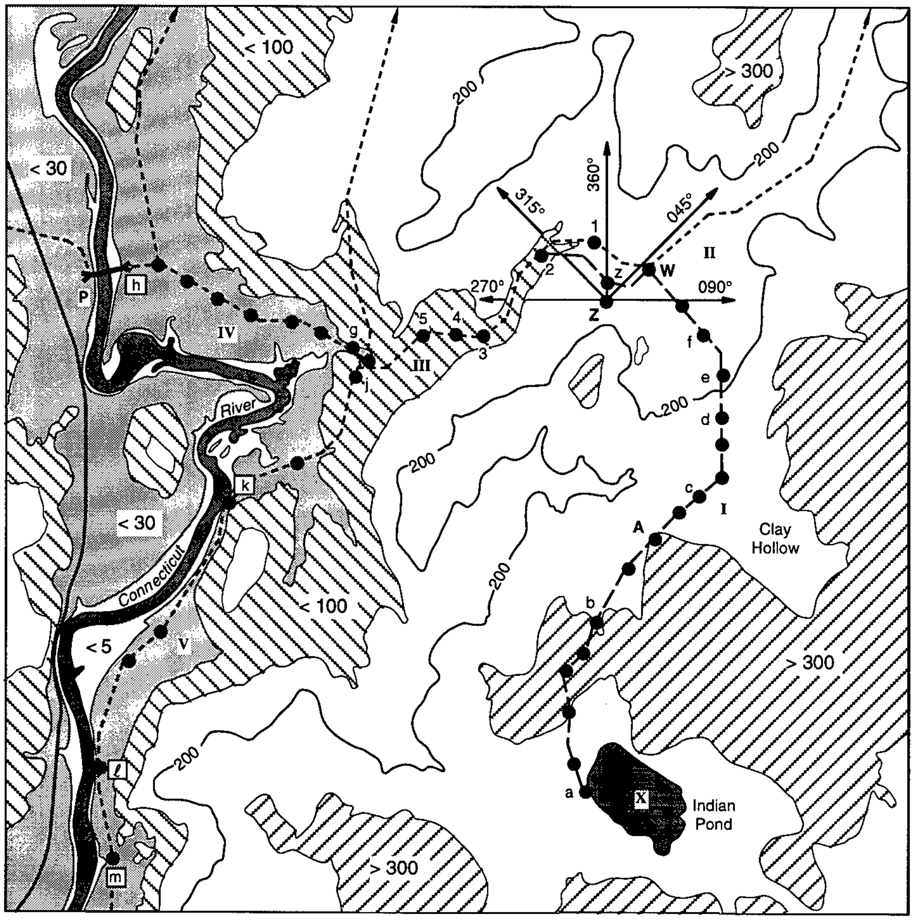


Figure 9. Enlarged map section showing topography (from USGS 7.5-minute quadrangle) of the area where large morning air temperature differences were observed. The symbols coordinate this figure with those in Parts I and III. The stippled layer lies 5 to 30 m above river level, the hatched layer 30 to 100 m above river level, and the alternate hatched layer 300 m above river level.

It is also necessary to critically examine the relationship among measured snow depth (Baker et al. 1991, 1992), observed snow cover, and inferred vertical temperature structure to support objective analysis of the influence of snow cover (Namias 1985) and local features on temperature difference. A quasi-simultaneous set of temperature observations were made on winter mornings in varying terrain along a curved path within 6 km of the Connecticut River, in the area east of the Piermont bridge along *P–W–A* in Figure 9. Figure 9 shows detail of this transect route, within the experiment area defined in Part I. The coordinated designators are again noted. Roman numerals are used to designate small basins in Figure 9; capital letters designate fixed observation sites, or important landmarks, along the measurement path cited in the text. Lower case letters and Arabic numerals are used to designate individual temperature observation points cited in the text, tables, or figures.

ANALYSIS OF VERTICAL TEMPERATURE STRUCTURE IN THE STUDY AREA

The temperature differences along the path *P–W–A*, with respect to the temperature (T_p) at the bridge *P*, were stratified to produce 10 quasi-homogeneous data subsets. The temperature difference *W–A* was again used to identify apparent lapse or inversion structure over the study area. The mean temperature difference at each observation point was calculated for each subset; the five subsets corresponding to lapse conditions are

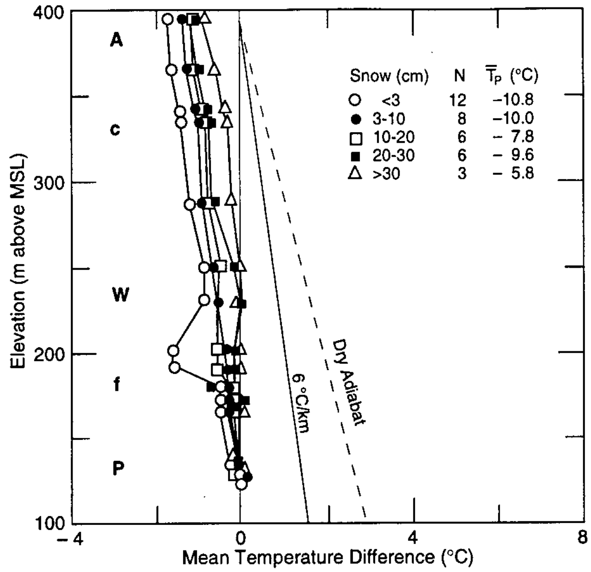


Figure 10. Mean elevation structure of surface air temperature, stratified by snow cover, when lapse was apparent along P–W–A. *The mean temperature at the reference point* P *is given in the reference block for each category.*

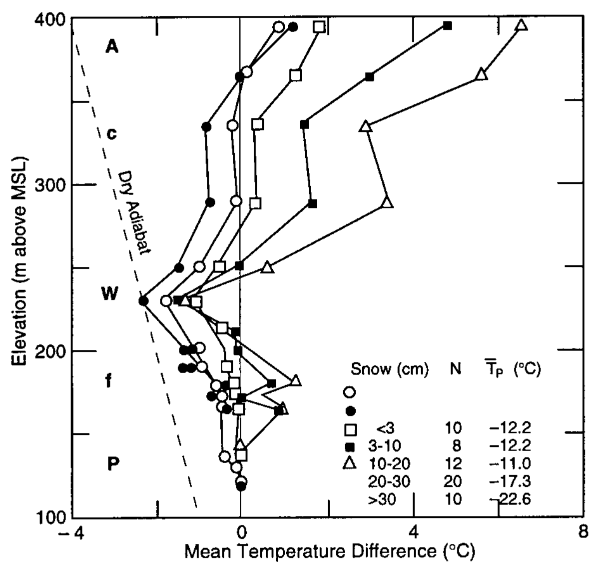


Figure 11. Mean elevation structure of surface air temperature, stratified by snow cover, when inversion was apparent along P–W–A.

plotted in Figure 10, and the five subsets representing inversion are plotted in Figure 11. The snow depths at Z defining the subsets and the mean reference temperature at P for each subset are noted in a data block in each figure.

A broken line representing the dry adiabatic lapse rate and a solid line representing a lapse rate of 6°C/km are included for reference in Figure 10, to guide interpretation of the temperature structure. (Isothermal or inversion layers may be embedded in the general structure.) When snow cover at Z was less than 10 cm and the cover throughout the valley discontinuous, a mean lapse of 6–10°C/km was present in the river basin below 180-m msl elevation. There was a mean tendency towards isothermal structure below 230 m, and a lesser overall (A–P) lapse structure as snow depth increased and became more uniform through the area. The inflection at 200-m elevation with minimum snow depth coincides with the broadening of the narrow stream valley and may be a cold air drain.

The mean temperature differences at the same elevations, calculated for inversion days when the temperature at A exceeded those at lower elevation, are shown in Figure 11. The lower river basin was characterized by lapse or isothermal structure beneath the mean inversion elevation when thin or discontinuous snow cover prevailed. There was inversion structure in the lower basin when snow was deeper and more uniform.

The analyses in Figures 10 and 11 support our proposal that the temperature difference between points A and W effectively defines the general lapse or inversion structure of the near surface air in the study area. The figures also indicate that thicker and more continuous snow cover was accompanied by isothermal structure in the 100 m of air just above the river plane when lapse was present at greater elevation. An additional surface-based inversion was present above the river plane when an inversion in the A–W layer was present over thicker and more continuous snow cover.

Compare the data blocks in Figures 10 and 11. The mean temperature observed at P increased with increasing snow depth under lapse structure. The mean temperature at P diminished as snow depth increased under inversion structure. Uniform snow cover along the river plane coincident with the elevation of P enhanced cooling and stratification in the largest, lowest basin during this period of study. Only well-organized (usually precipitating) low pressure systems were able to displace this stable air in early morning hours. The relatively few occurrences of lapse temperature structure over thick uniform snow cover accompanied or followed precipitation. There were many occurrences of precipitation falling through inversion temperature structure in one or more layers.

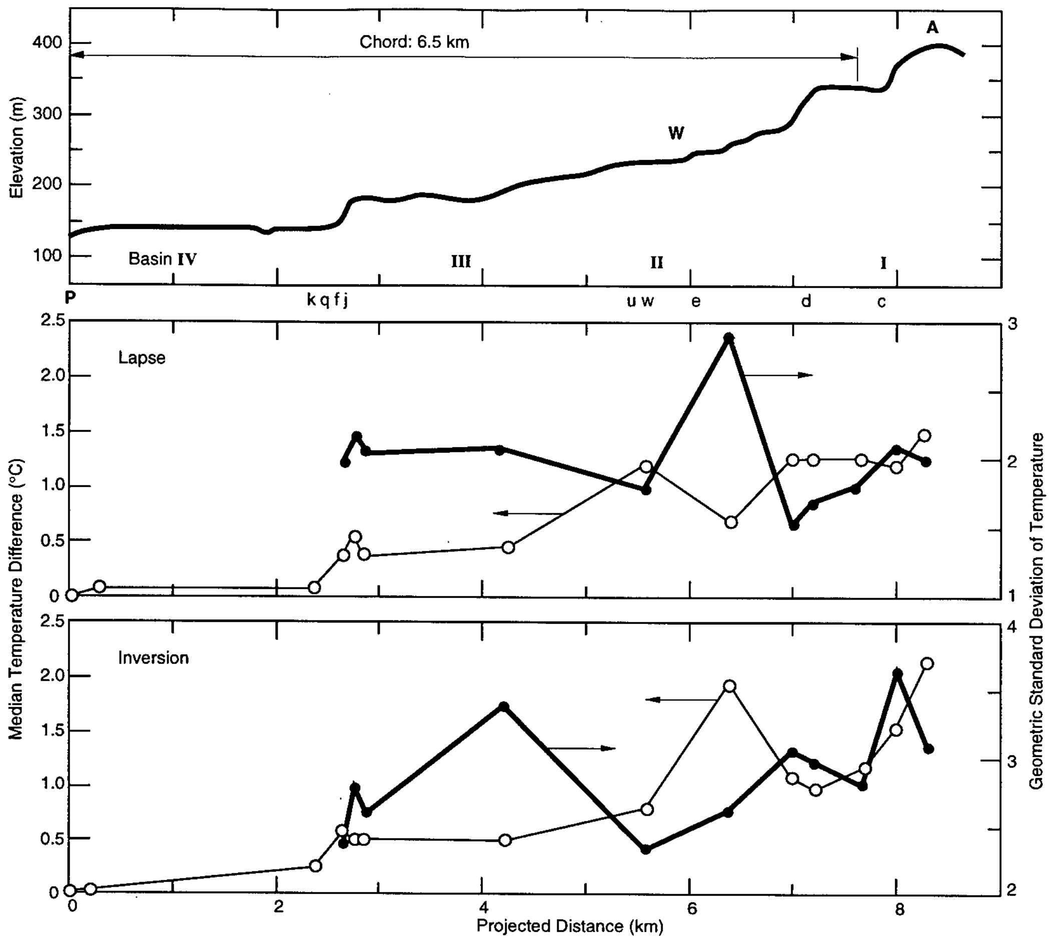


Figure 12. Median temperature difference and geometric standard deviation of temperature difference along the path P–W–A. *An exaggerated topographic cross section shows the slope in the vicinity of each measuring point.*

An exaggerated elevation cross section along the observation path *P–W–A* of Figure 9 is shown in Figures 12 and 13. The distance between measuring points has been conserved in the elevation projection of Figures 12 and 13. The coordinated designators may be used to identify the points on the figures, the maps, or in Parts I and III. Rhumb line distances between nonadjacent observation points are best estimated from Figure 9.

The median value of the absolute temperature difference, with respect to the temperature observed at bridge *P*, at observation points along *P–W–A*, is plotted for lapse and inversion conditions in Figure 12. The geometric standard deviation (GSD) of the temperature differences are also plotted. Note that the scales of GSD differ; systematically greater GSD occurs with inversion structure. The median temperature difference with respect to *P* increases with elevation as expected under lapse, but an inflection in both temperature difference and GSD occurs at *W*. The inflections reverse under inversion. The largest absolute temperature difference occurs at *A*, the apex of observations, under both lapse and inversion.

The 84th percentile of temperature difference, with respect to *P*, is shown in Figure 13. The temperature differences shown are again the absolute difference in temperature with respect to that at the bridge *P*, as in the temperature differences along the river shown in Part I. The frequency of lesser temperature at the point, with respect to *P* (i.e., the fraction of times that it is colder at the point than at the bottom of the valley) is also shown for lapse and inversion. The temperature at *A* is not uniquely lesser than at *P* under lapse conditions, as we have stratified our data on the temperature difference in the layer *W–A*. Under inversion conditions the air temperature at elevations near that of *W* are frequently less than that at *P*, and the frequency of lesser

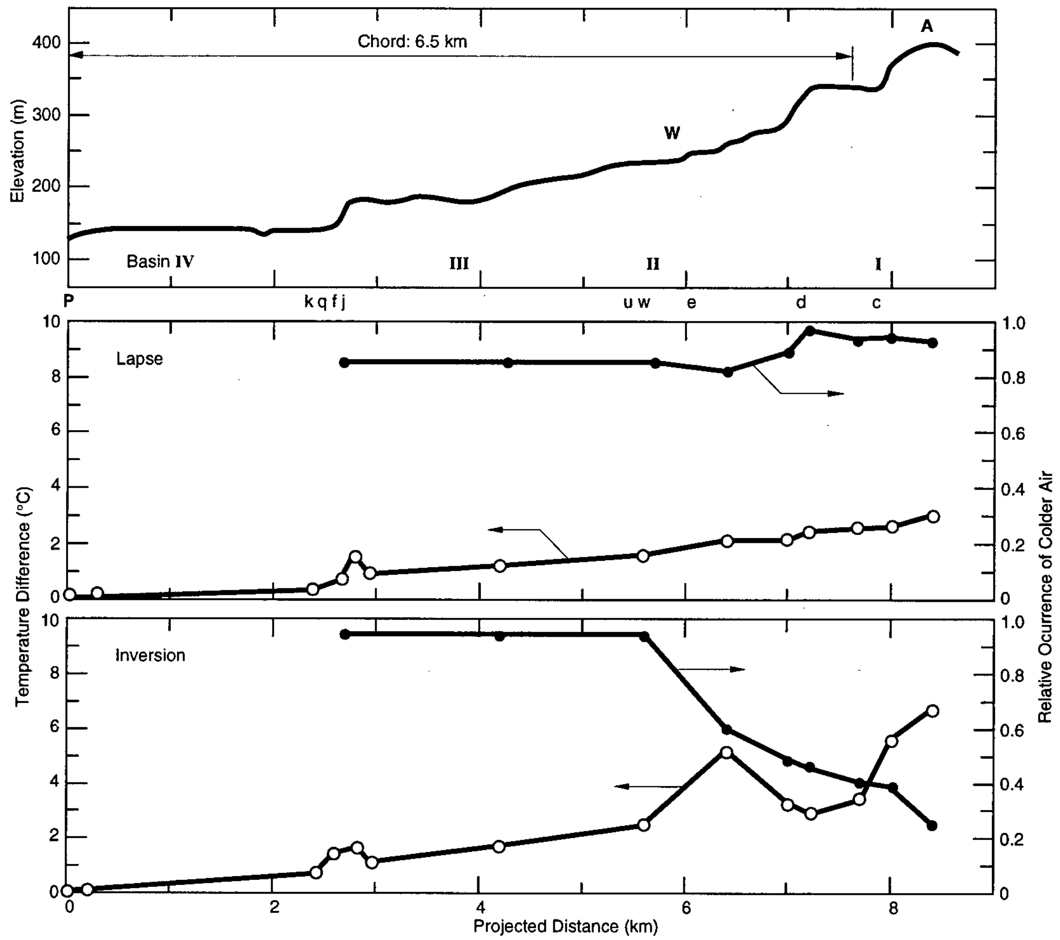


Figure 13. Eighty-fourth percentile of temperature difference, and frequency of lesser temperature with respect to P, *along the path* P–W–A.

temperature diminishes as elevation increases above the *W* level.

The analyses in Figures 12 and 13 indicate the frequency of greater temperature difference, and the geometric standard deviation of temperature difference with respect to the lowest point in the valley (*P*), increase at elevations greater than 230 m. The frequency of lesser temperatures with respect to *P* diminishes above this elevation under inversion. This is the most frequent elevation of the inversion base in winter.

Recombining the data subsets of Figures 10 and 11 shows the coldest mean air temperature occurred near *W* in basin II at 230 m. This is coincident with the minimum value of GSD of temperature difference in Figure 12 and the inflection in frequency of colder air temperature in Figure 13. These statistical arguments can be summarized with the statement that air below 230-m elevation is least influenced by advective warming on winter mornings. The layers at or below 230 m are decoupled from tropospheric advection on many winter mornings, allowing the Connecticut River Valley to experience colder air than the summit of Mt. Washington.

Air temperature structure above this persistent inversion base was dependent on the continuity of snow cover during the observation period, as shown in Figure 11. A point (*c* in Fig. 9) in the Clay Hollow basin (*I*) has a lesser air temperature than points along nearby slopes, and it has a lesser temperature than the lowest point *P* on 0.4 of inversion days. This small basin had the coldest air temperature in the study area on several days when the apex of observations (*A*), only 0.6 km distant, had the warmest air temperature. The basin-slope temperature differences observed along *P–W–A* are similar to those found near Moshiri by Nakamura and Magono (1982), Magono et al. (1982), Maki and Harimaya (1988), and Maki et al. (1986).

The same stability and snow cover characterizations were applied to stratify air temperature

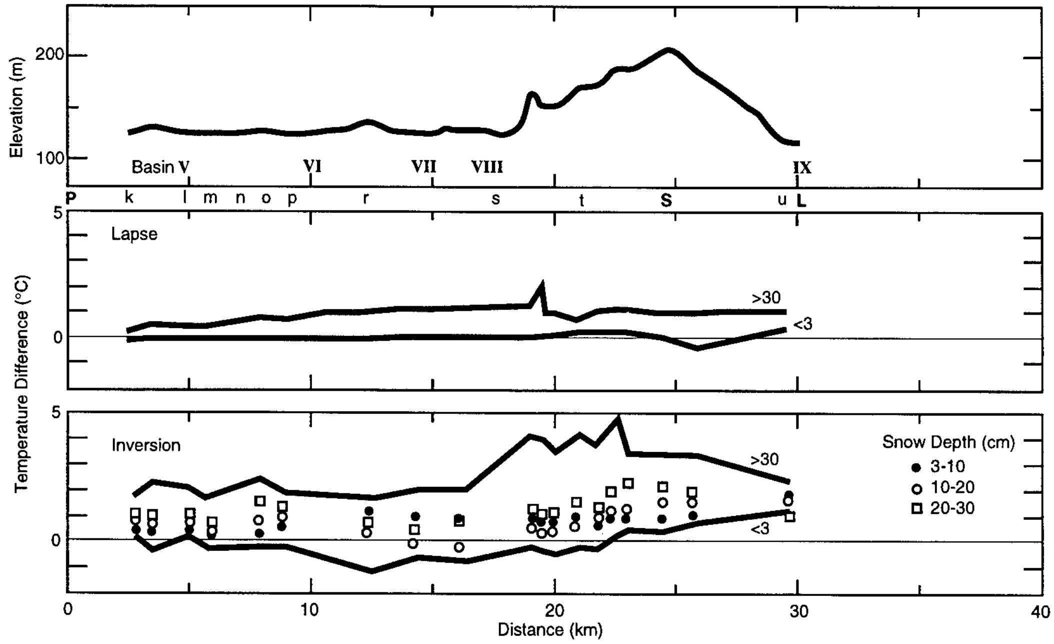


Figure 14. Mean north–south cross sections of mean surface air temperature, with a terrain cross section for reference. The morning air temperatures are stratified by snow depth and elevation structure. The envelopes associated with maximum and minimum snow depth are plotted to guide the eye.

difference subsets observed along the 30 km north–south observation path from *P* to *L* discussed in Part I. The mean temperature differences, with respect to the temperature at *P*, were calculated for these data subsets. Cross-sectional analyses of the mean temperature differences, including the near river points for which frequency distributions were given in Part I, are shown in Figure 14. Coordinating symbols are again noted on the topographic cross section.

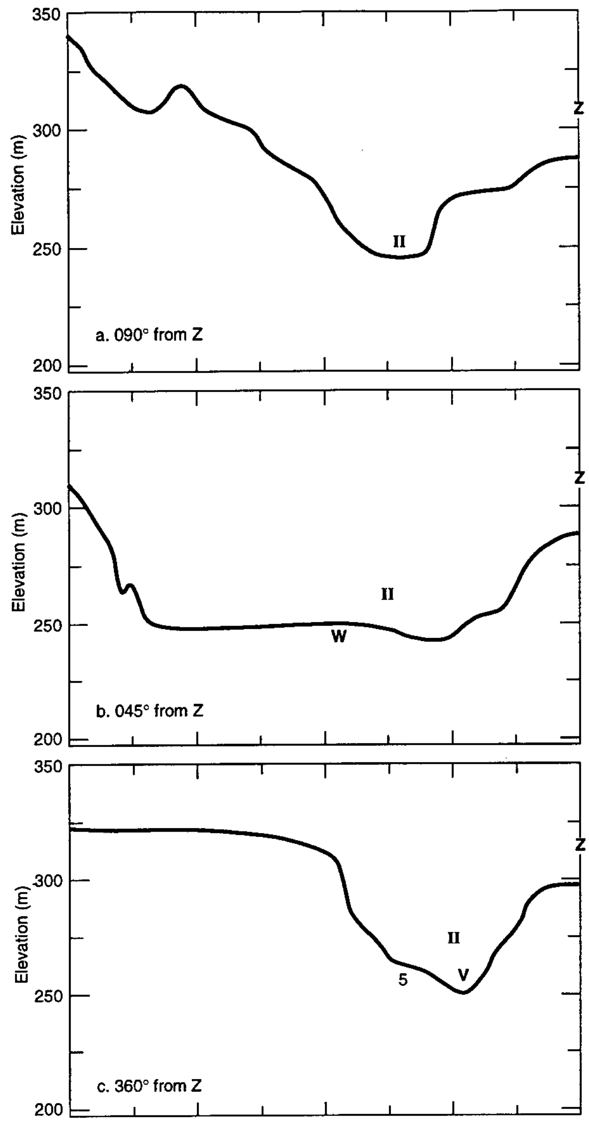
Under lapse structure, air temperatures along this transect are related to the temperature at the bridge *P* by the lapse rate. Because of the narrow range of temperature differences, only those measured with sparse snow cover (<3 cm) and complete snow cover (>30 cm) are plotted. These points are connected to form an envelope to guide the reader. All other temperature differences under lapse are enclosed in the plotted envelope. This small temperature difference necessitated use of mean values in this analysis.

The temperature differences at these places are much greater under inversion. All mean morning temperature differences corresponding to the snow cover categories are plotted at each observation point. The envelope lines of the sparse and total snow cover classes are again plotted to aid in estimating the range of variation.

We established, in Part I, that the temperature difference along the river plane was less than 1.6°C over 30 km in 84% of observations. Figure 14 indicates that several small basins at slightly higher elevation had morning air temperature differences quite similar to points along the river plane. Temperature difference along the slope south of *S* was greater than that in those basins (V, VI, VII, and VIII). This difference is similar to the slope basin temperature difference on the slopes connecting basins II, I, and the apex of observation, *A*. These temperature differences generally conform to the slope and basin theories of Magono et al. (1982) and Maki et al. (1986). It does appear that depth and continuity of snow cover may be additional variables influencing the morning temperature on slopes and in basins.

EXAMINATION OF SOME SINGULAR EVENTS

The experiments, observations, and theories describing winter slope and basin temperatures in Hokkaido were primarily formulated to explain the extremely cold air that occurred on a few days. Hokkaido has very volcanic topography, and most of the work of Magono et al. (1982)



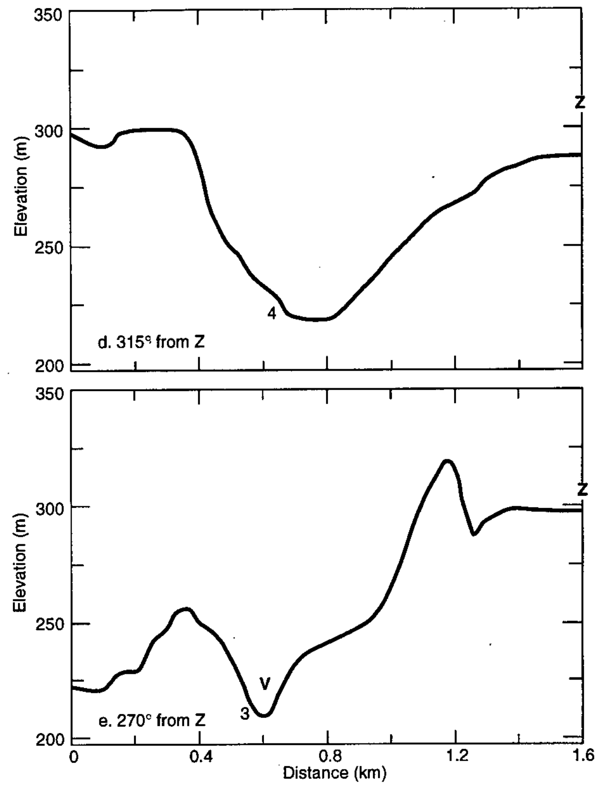


Figure 15. Topographic cross sections at 45° direction intervals extending radially from Z. Landmarks and observation points from Figure 9 are provided for orientation.

and Maki et al. (1986) was accomplished when 1.5 m or more of snow covered the ground. Our experimental area has many slopes and basins, but local relief is generally less than that of Hokkaido. The Connecticut River Valley observations are at lower elevation but similar latitude to the Hokkaido observations, and include snow depths of 0 to 0.75 m. The minimum temperatures observed in the two locales were quite similar during the experiment periods.

The data of Figures 10 through 14 indicate that, under inversion, the coldest mean morning air temperature and the greatest frequency of coldest morning air temperature in the area occurred in the small basin (II) surrounding *W* in Figure 9. When the coldest air temperature did not occur at *W*, it most frequently occurred at the bridge *P* when continuous snow of >20 cm was present, or at *c* in the small basin (*I*) at Clay Hollow at 340-m elevation. There were fewer than five singular events where the coldest morning air temperature was observed at another place. We thoroughly examined, in Part I, the relation of air temperature at *P* to other places in the Connecticut River Valley. A rigorous examination of the temperature structure near *W* and *c* is necessary to assess the processes which may diminish air temperature at those places.

Although Figure 9 is the largest scale map available, it does not clearly establish that *W* is a basin that narrows to a small valley as one proceeds downslope. Topographic cross sections extending radially from *Z*, in Figure 9, at 45° intervals are shown in Figure 15. The upper entrance to the

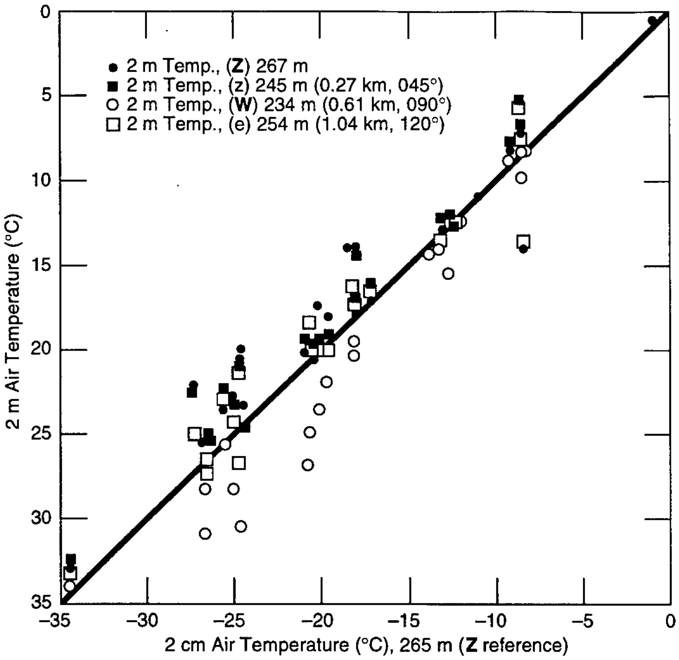


Figure 16. Temperatures in and around basin II compared with the 2-cm temperature at Z. The temperature in basin II is consistently less than the 2-cm temperature on the slope above.

basin is shown in the east (090°) radial section; the northeast radial section (045°) shows the broadening of the basin at *W*, and the north (360°) radial the narrowing of the basin at the entrance to the valley. The narrow valley is shown in the 315° radial, and the valley begins to broaden into a basin (III) in the 270° radial section. The nature of Basin II surrounding *W* is not readily apparent in a west–east elevation section, but the bounded basin is quite apparent in the cross section along the 045° radial in Figure 15.

The basin (II) surrounding *W* may receive air draining from the adjacent slopes (Yasuda et al. 1986) and retains it through boundary damming by the more wooded terrain (Baines 1979, Raynor 1971) along the valley (Yasuda et al. 1986) which drains the basin on the west. A thin stagnant layer in the bottom of the basin could then contain cooler air than found nearby. A test of this premise was performed during winter 1993. A series of temperature measurements, 2 cm and 2 m above the snow, were obtained at the fixed site Z to compare near snow temperature along the slope to the air temperature in the basin. The temperature measured at 2 cm at point Z during January–February 1993 is defined as the reference, and plotted as the abscissa in Figure 16. The corresponding 2-m air temperature at Z, the 1.25-m air temperature along the 045° radial between Z and *W*, the 1.25-m temperature at *W*, and the 1.25-m temperature observed at *e* (projected to the 090° radial in Fig. 15) are plotted against this 2-cm temperature in Figure 16. The elevation of each measuring point and its relative direction from Z are tabulated in the data block in Figure 16. The 1.25-m air temperature in the basin (*W*) is most frequently less than the temperatures around the basin, and is 1 to 5°C less than the 2-cm temperature above the basin in many cases.

The observation path was continued beyond the apex of observation *A* to the Indian Pond basin (X) at *a* on several mornings. The observations along *A–a* were used in conjunction with the observations along the usual path to construct a temperature–topographic cross-sectional analysis. This enlarged elevation section is shown in Figure 17. The temperatures observed at 0630 on 7 February 1993, the coldest morning of the three winters (1990–1993), are plotted on the cross section in Figure 17, and as one of the significant vertical profiles of air temperature in the Connecticut Valley, in Figure 18a.

Very strong, cold advection with surface wind of 5–10 m/s was present throughout the area on 6 February 1993. At 2130 EST, the surface winds had diminished, and the temperature at *W* was –30.8°C, this being the lesser of several air temperatures measured in the vicinity that evening. Warm advection increased the temperature at Mt. Washington to –20°C by 0630 on 7 February, but radiational cooling and inversion strengthening continued in the study area. Air temperature below the ridge tops was consistently less than –30°C during the 0615–0700 period of observation. The points *A–a* on the posterior of the ridge from *A* as in Figure 17 are shown by open circles in Figure 18a. The coldest air was in basins (I) and (II), evident in both Figure 17 and Figure 18a. Lapse structure was present from river level to 230 m although considerable temperature differences occurred at each elevation. Examining the cross section of these air temperatures with respect to terrain in Figure 17 indicates that the isotherms quite consistently intersect the ridges, and that the Clay Hollow and Indian Pond sides of the barrier ridge have similar air temperature structure. The inversion isotherms above Clay Hollow and Indian Pond are quite closely packed,

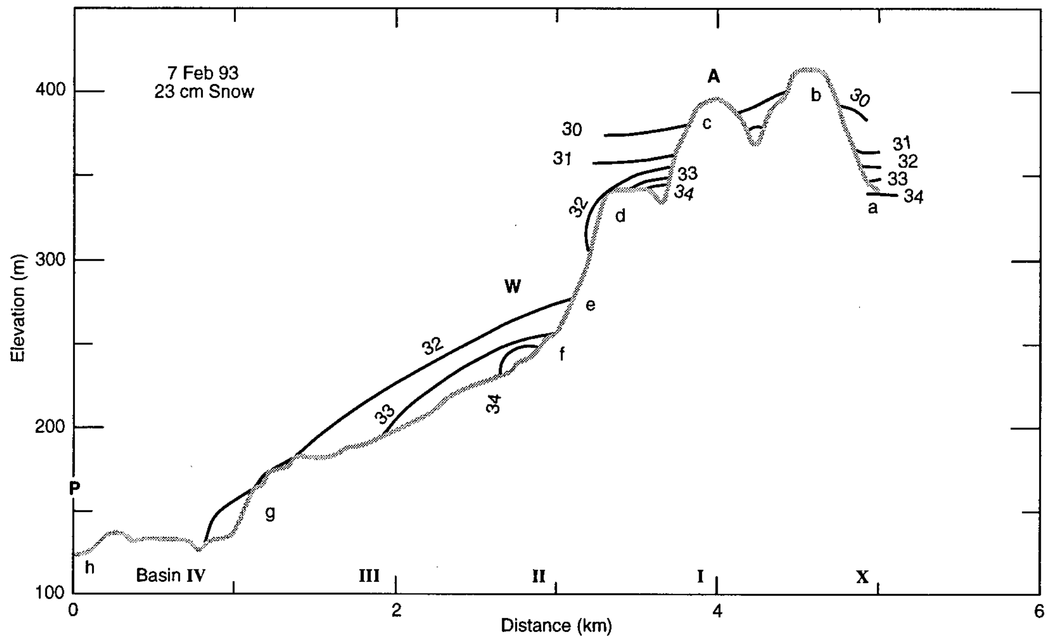


Figure 17. Temperature–topographic cross section obtained 0700LCL 07 February 1993, the coldest air observed during the winters 1990–93. The isotherms are inferred from intersection with the 1.25 m above ground level surface.

separated by only 5 to 20 m per °C. The analyses in Figures 17 and 18 indicate that the layers of coldest air above the small basins are at most a few tens of meters deep.

Additional vertical profiles for comparison with this day are given in Figure 18, illustrating some extremes of temperature structure found near the Connecticut River. Nocturnal cooling following cold advection is again shown in the vertical temperature profile of 2 February 1993 (Fig. 18b). This is one of many days when an inversion formed above *W*, while most other observed temperatures diminished with increasing elevation.

The coldest "snowless" day of observation occurred 1 January 1991 (Fig. 18c), with grass extending through less than 3 cm of a discontinuous fresh dusting of snow. Another day with rather shallow snow, 10 December 1992, is shown in Figure 18d. Significant radiational cooling (sufficient to reduce surface air temperature to less than the frost point) occurred, producing large frost crystals atop 5 cm of snow.

The changes in vertical temperature structure associated with cold and warm advection are shown in the profiles of 21 through 23 January 1991 in Figures 18e and f. Vigorous cold advection caused temperature diminution of 4° to 6°C between 2100E21Jan and 0600E22Jan, noted on the 22 January profile. There is lapse structure in the lower layers, a small inversion over basin (II) surrounding *W*, and near-isothermal structure in the *W–A* layers. The cold advection weakened through 22 January, and a profile observed at 1700E showed inversion structure forming above the elevation of *W*. A repeated observation at 2200E22Jan showed inversions over both basins (I and II), and 4° to 9°C temperature diminution since 1700E. Strong warm advection began after 2200E, which caused warming of 1°C at *W*, 3°C at Clay Hollow, and 3° to 6°C along the ridge about the apex. This provided 7°C of inversion (150°C/km!) over basin *I* in Clay Hollow and a mean inversion of 50°C/km in the layer *W–A*, while lapse structure remained in the Connecticut River Basin below *W*. Magono et al. (1982) proposed that the initial inversions in an area formed over snow-covered basins, decoupling the air in the basins from exchange with warmer tropospheric air above for longer periods than adjacent slopes. This decoupling persisted for several hours after the onset of warm advection, in the case shown in Figure 18f.

Each panel of Figure 18 shows a layer of air, of about 15-m thickness, in contact with the surface of the basin surrounding *W*, which is persistently of lesser temperature than nearby air. This smaller basin appears to respond, over lesser snow cover,

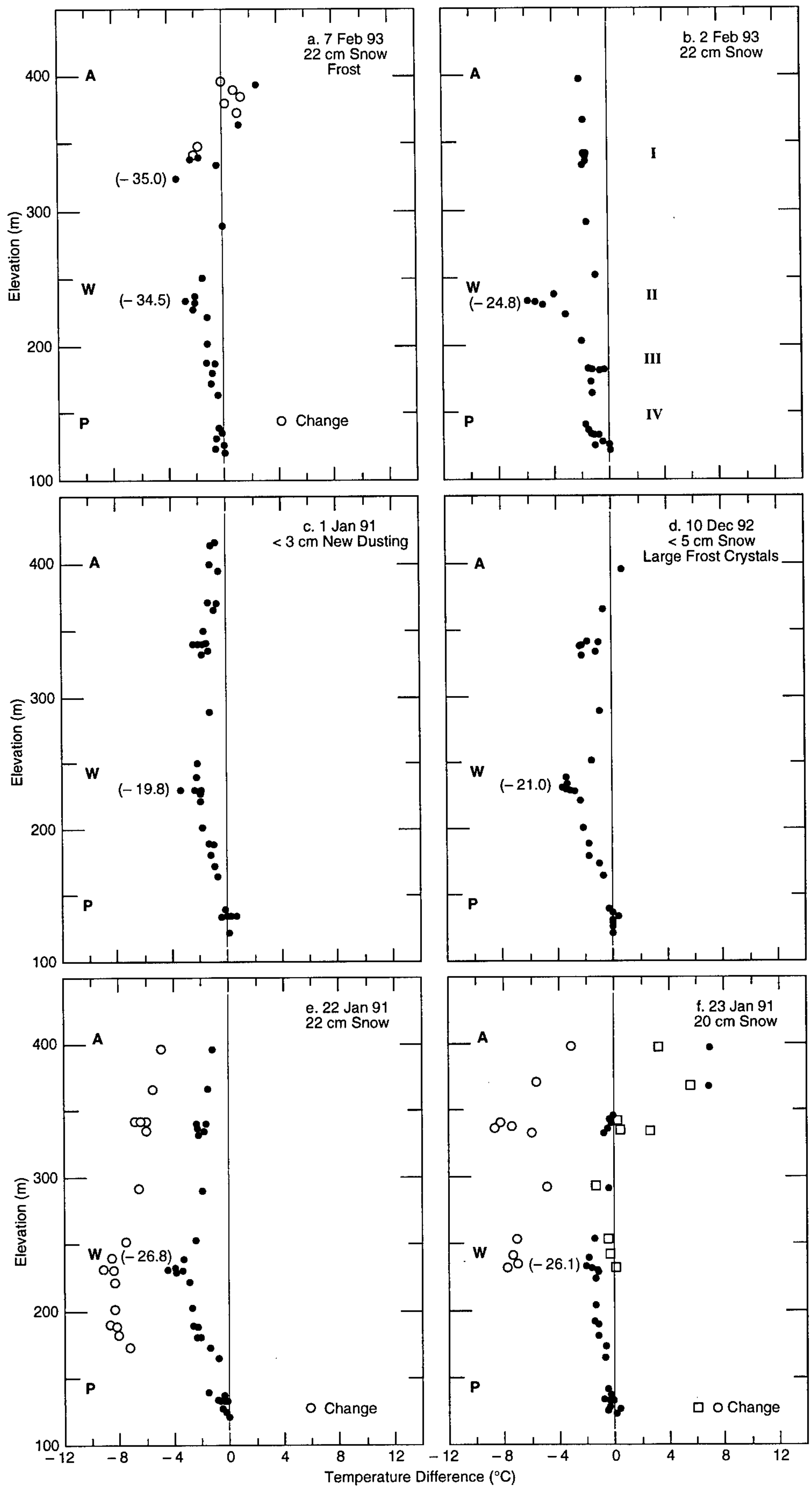


Figure 18. Vertical profiles of 0700LCL air temperature extracted from surface observations along P–W–A: a) The coldest day of the study period, 7 February 1993, following strong cold advection throughout 6 February. The open circles indicate temperature change since 2100 6 February. b) Temperature profile over 20 cm of snow 2 February 1993, following 24 hours of strong cold advection. c) The coldest snowless day, 1 January 1991; grass protruded through a dusting of less than 3 cm of new snow. d) A day with significant radiational cooling, 10 December 1992; large frost crystals formed on <5 cm of snow. e) Following cold advection 22 January 1991; the open circles indicate temperature change at that point since 2100 21 January, over 20 cm of snow. f) During warm advection 23 January 1991; the hollow squares indicate temperature change 1700–2200 22 January, the hollow circles indicate temperature change 2200 22 January to 0700 23 January over 20 cm of snow.

in the same manner as the larger, more steeply sided basins with greater snow cover, described by Nakamura and Magono (1982), Magono et al. (1982), Maki et al. (1986), and Maki and Harimaya (1988). This is consistent with the theories that early formation of inversion and diminished exchange of tropospheric heat from greater elevations result in diminished temperatures in snow-covered basins. The similarity of isotherm intersections at Clay Hollow in Figure 17 is indicative that the same mechanism causes the frequently diminished air temperatures observed there.

SOME EXTREME TEMPERATURES OBSERVED COINCIDENT WITH UNIFORM SNOW COVER

It may be significant that the largest and lowest basin in the study area, along the Connecticut River, rarely attains the coldest local morning air temperature under snow cover of less than 20 cm. The temperature differences observed in replicate measurements under inversion conditions in Part I may reflect exchange of warmer air from aloft to the layer within 10 m of the surface, as proposed by Magono et al. (1982). Snow cover of greater than 20 cm is accompanied by more frequent occurrences of the coldest air at *P* in the river basin. Deeper snow cover may provide a more Hokkaido-like, uniform snow surface in this lower basin, promote inversion formation, limit exchange from aloft, and allow colder surface air temperatures there.

The vertical profiles in Figure 19 represent days from the winter of 1993–94, which were quite unusual. These observations are comparable to the 1990–93 data set as the same instrumentation and measurement protocols were used. The time period 20–27 January 1994 contains the authors' only verifiable observations of air temperature less than –35°C at less than 45°N latitude and lower than 1-km elevation in 30 years of experience. The same temperature difference scale is used in Figures 18 and 19 to allow comparison.

The ground was covered by 60 cm of snow on 20 January 1994 (Fig. 19a) and strong cold advection conveyed air of –33°C at 0700EST to Mt. Washington. There was no temperature mea-

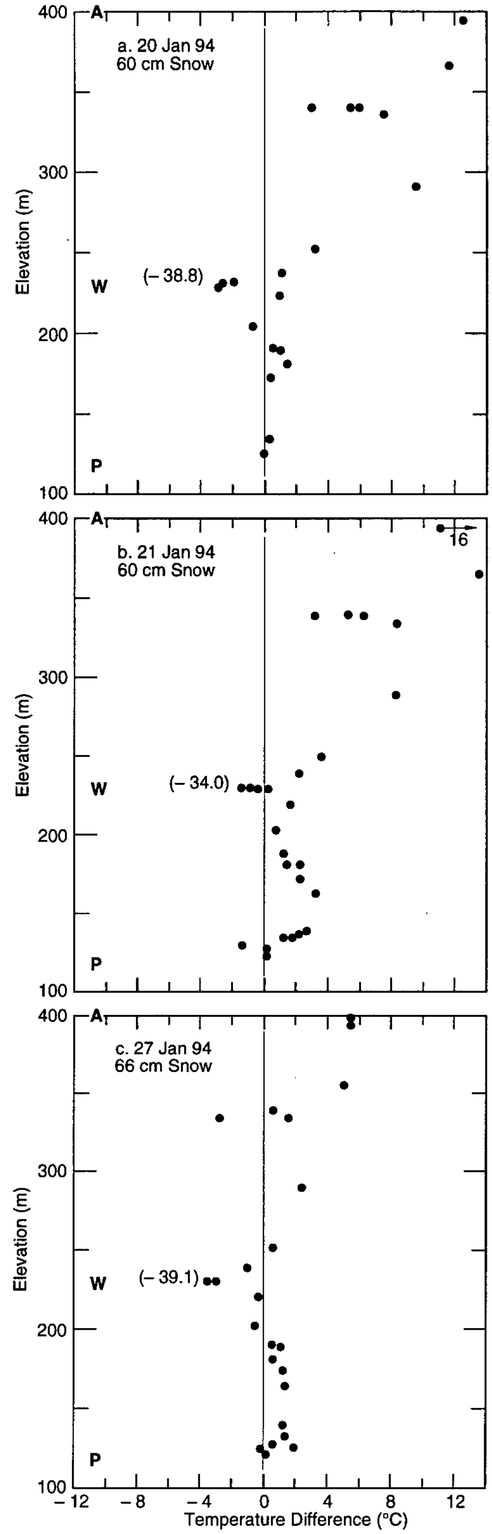


Figure 19. Temperatures along the same observation path, observed with the same instrumentation, during a very cold period in 1994: a) An inversion is present from 125 to 220, which strengthens above 230 m, on 20 January 1994 over 60 cm of snow. b) Strong inversions above the river and in basins I and II 21 January over 60 cm of snow. c) An extremely cold day, 27 January 1994, with suspended ice crystals, 66 cm of snow, and inversions over the river, basins I and II. Warm advection had begun at the time of this observation at Mt. Washington.

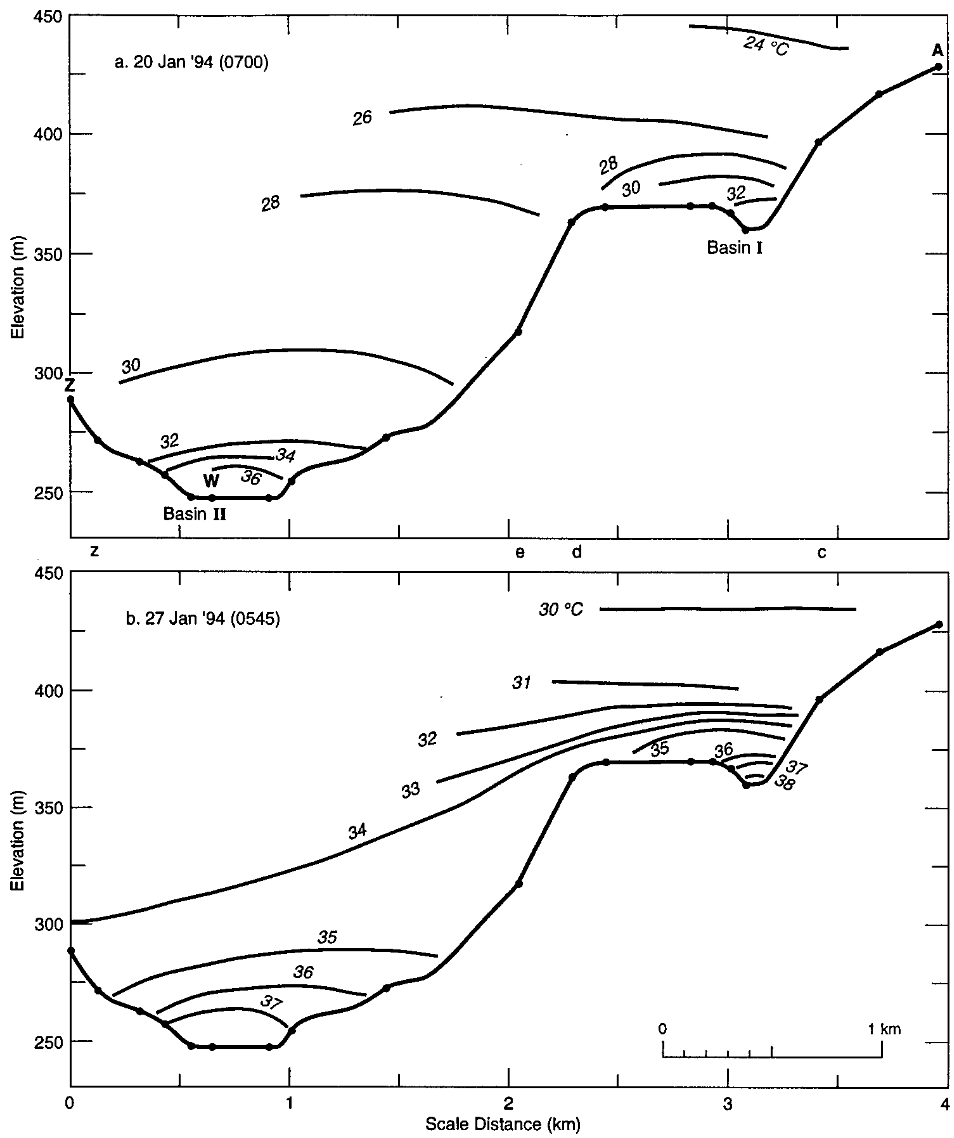


Figure 20. Inferred isotherms above the topographic cross section along Z–W–A *on 20 (a) and 27 (b) January 1994, when very cold air was observed.*

surement on bridge *P* on 20 January, but measurements at *k, l, m* near river level indicate isothermal or inversion structure was present in the river basin. There was 5°C of temperature difference within the basin surrounding *W* and in Clay Hollow. The apex (*A*) was 5° to 10°C warmer than Clay Hollow 0.6 km distant, 11°C warmer than *W*, and 5°C warmer than Mt. Washington.

Warm advection occurred during the night of 20–21 January, producing an extreme inversion. Measurement at *P* on 21 (Fig. 19b) January showed inversion structure in the lower basin. The air was about 5°C warmer in basins I and II than on the previous night, but inversion was very strong with a 10° to 13°C temperature difference between Clay Hollow observations and those near the apex, and a difference of 16° to 18°C between *W* and those near the apex.

Strong advection of arctic air occurred again on 26 January 1994, but warm advection had begun at Mt. Washington at 0700EST 27 January, where –16°C was observed. There were 70 cm of snow at *Z* at this time. A cross section was obtained on 27 January (Fig. 20b), and the vertical profile of temperature repeated the inversion profile shown in Figure 11. A somewhat similar advection event occurred 10 February 1994 but did not produce air temperatures less than –35°C.

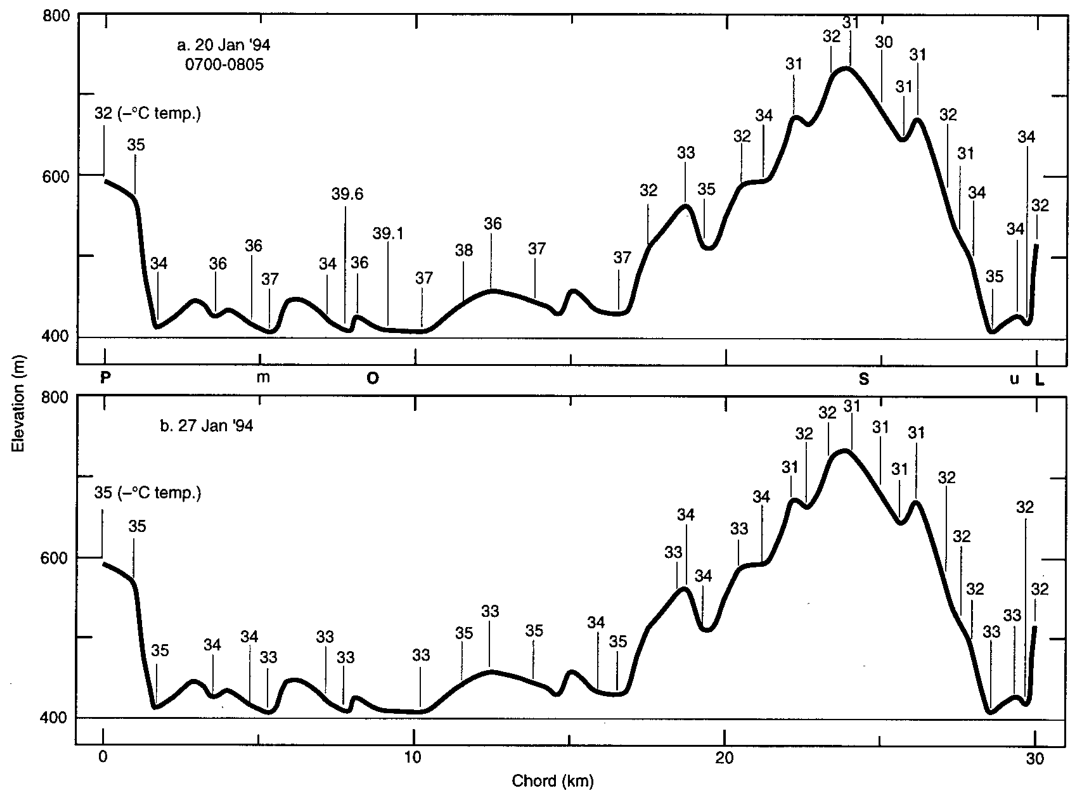


Figure 21. Temperatures observed relative to the topographic cross section along P–S–L *on 20 and 27 January 1994. Thin ice fog akin to the "clear sky precipitation" observed at the South Pole occurred along* m and north of O *on 27 January.*

Temperature–terrain cross sections of morning surface air temperature measured on 20 and 27 January 1994 are shown coincident with terrain features along the *Z-W-A* path in Figure 20 and the *P-S-L* path in Figure 21. The minimum temperatures observed at W (since 1988) occurred on these days, –38.9°C on 20 January and –39.1°C on 27 January 1994. The temperature gradients surrounding basins I and II were similar to those of other snow-covered periods shown in Figure 3. Basins I and II had lesser temperatures than Mt. Washington on both of these mornings, and significant warming had already occurred at Mt. Washington 27 January. It is important to note that on 20 January, air temperature of –39.6°C was observed concurrent with ice fog along flats at *m–O* along the lower river basin. This was the only observation of record where the coldest air of a morning was measured along this segment of the observation path, and it occurred under a very low (<15-m) surface inversion indicated by the flattening of smoke plumes of house chimneys. The general slope warmer-basin colder relationship remains on these very cold mornings, and the temperatures at points used in the river plane analysis of Part I differ by less than 2°C.

Cold advection continued to produce morning temperatures of –30°C through mid-February 1994, and "exploration" was conducted on several mornings to determine if some other nearby basins experienced lesser temperatures than those studied. Terrain–temperature cross sections observed 2 and 10 February 1994 are given in Figure 22, in similar format to Figure 21. These cross sections were obtained along a path from *Z* extending across basin II and then along a northeasterly radial following the road shown in Figure 9, passing through some additional small and large basins. The temperatures observed northeast of basin II are comparable to those south of the basin, and no colder air was found along this cross section on these or other days of exploration. The temperatures observed at several elevations along this transect are quite similar to those observed along the usual observation path.

Some similar exploration was conducted in the winter of 1991–92, concurrent with the extension of the river level analysis to bridge *H*, described

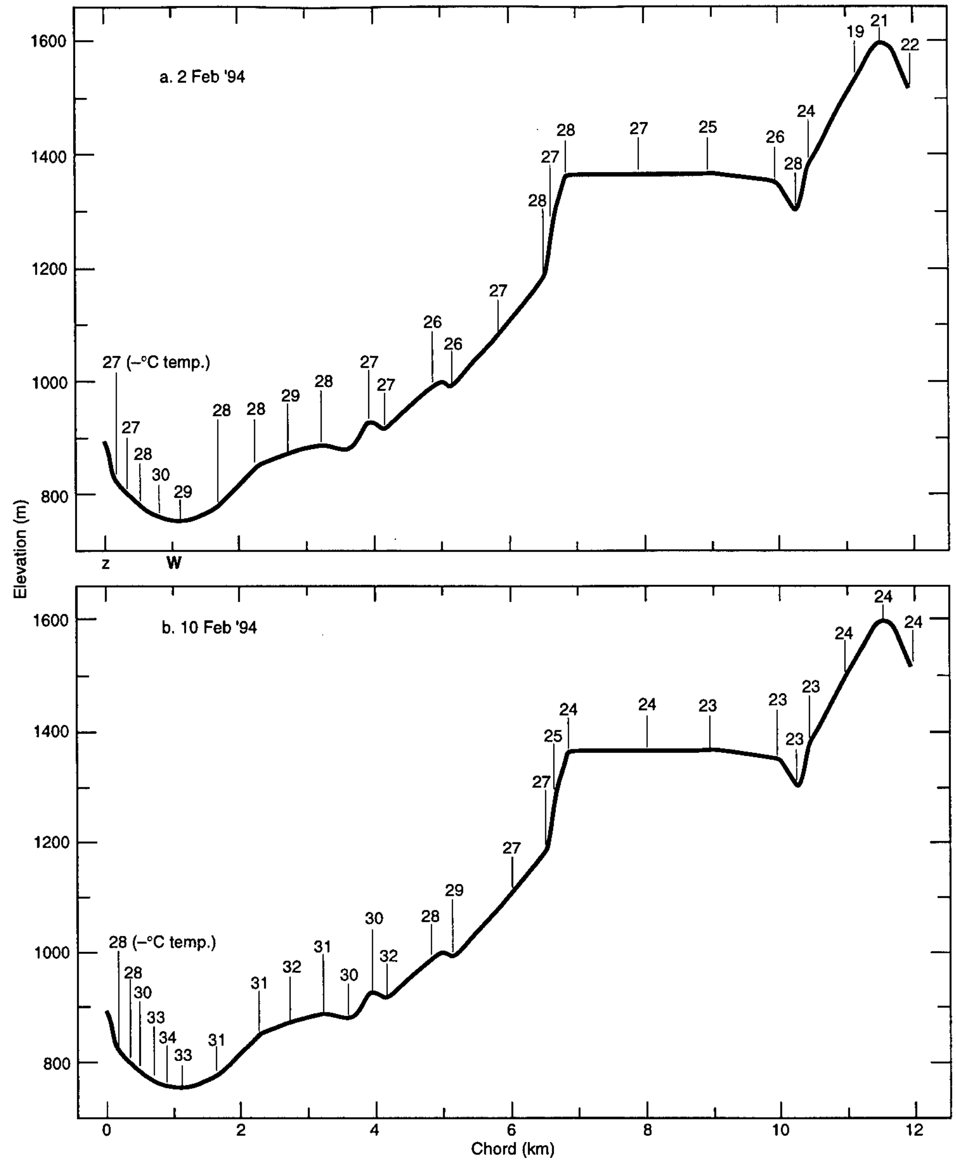


Figure 22. Temperature observed relative to the topographic cross section along the 045 radial from Z, *to a more distant ridge of greater elevation than* A. *Temperatures along other ridges were comparable to* A, *but other basins of similar size to basin II along Eastman Brook had greater air temperatures.*

in Part I. Temperatures were observed along several paths from *W* to *P*, *W* to *H*, and *H* to *P*. Temperatures observed over 10 cm of snow cover on 13 February 1992 are presented in Figure 23, as a function of elevation. The inversion which intersects the terrain slightly above basin II appears to intersect terrain at several places of similar elevation north of the usual observation path. While there are only a few mornings of exploration data available, these observations support the more general observations that an inversion above the Connecticut River intersects the terrain at 230 m, decoupling the layers below from warmer air advected over the study area.

The north–south observation path *P–S–L* reaches an elevation of 220 m at *S*, and returns to river level at *u*, 29.4 km south of *P*. Vertical temperature profiles extracted from observations along *t–S–u* on days with inversion along *P–W–A* are plotted in Figure 24. These vertical profiles of temperature difference with respect to that at *P* are stratified by snow cover class as in Figures 10 and 11, but contain less days of observation. Lapse or isothermal structure predominates when snow

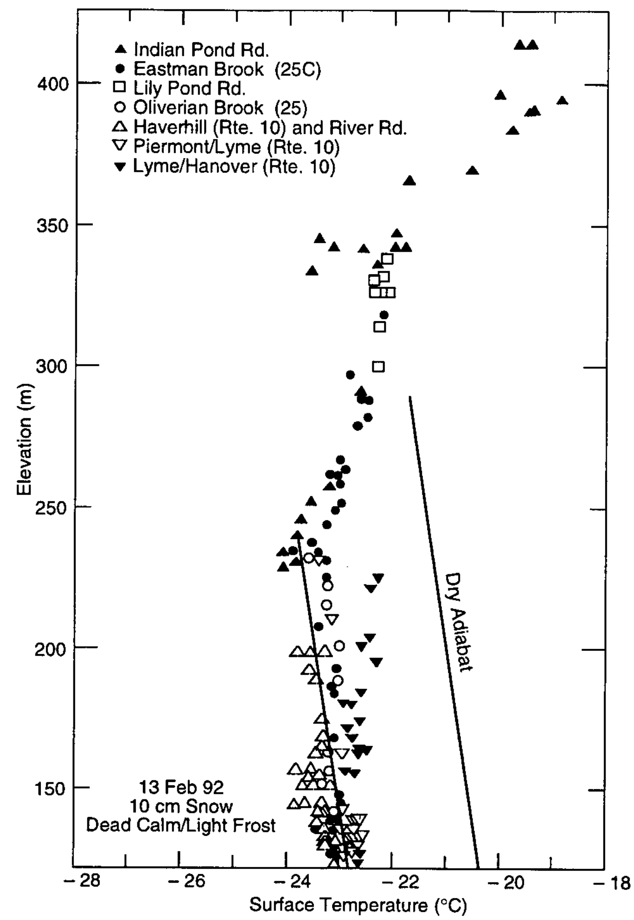


Figure 23. A composite plot of elevation and surface air temperature at that elevation observed along several paths within 10 km of the usual observation path on 13 February 1992.

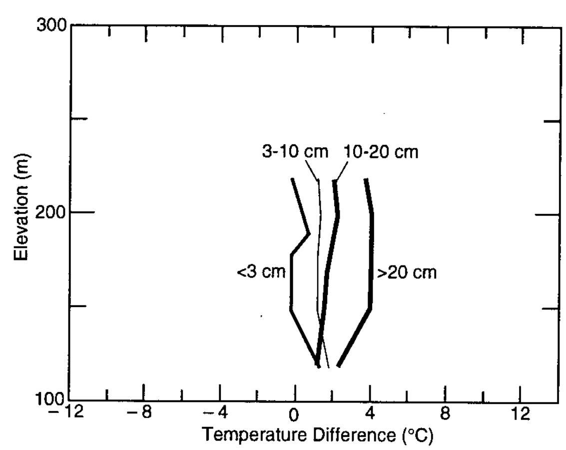


Figure 24. Mean vertical temperature structure constructed from surface observations north and south of S in the same manner as Figure 11. The database includes about two-thirds of the inversion days used to construct Figure 11.

cover is sparse. Inversion structure becomes more frequent as snow cover deepens and becomes more continuous. The inversion based at the plane coincident with the river bank becomes prevalent throughout the study area when thicker more continuous snow cover is present.

We proposed, in the introduction to Part I, that vertical temperature structure around the Connecticut River could be inferred from a field of surface observations along terrain of varying elevation. The temperature difference at two elevations was used to stratify observation mornings in apparent "lapse" and "inversion" categories, and yielded time and space variation of surface temperature consistent with those stability categories. We showed, in Figures 10 and 11, that the temperatures over small basins, along slopes, and in the basin along the Connecticut River responded consistently to the inferred lapse of inversion structure. We then showed through frequency analysis that the temperature inflections tending toward minima, of Figures 10 and 11, coincided with basins in topographic cross section. We verified through analysis of individual days of varying snow cover and synoptic conditions that the temperature inflections over slopes and basins were persistent features, and that these basins did indeed begin cooling earlier at night than surrounding terrain. These observations are all consistent with those of Magono et al. (1988) and Nakamura and Magono (1982), who measured similar winter temperature structure with an array of fixed thermometers and tethersondes.

We have not measured the vertical temperature profile in the lower 300 m of air over the Connecticut River on any observation day. All of our results are consistent with our inference that measurement of surface temperatures at two elevations sufficiently defines lapse or inversion for analytical purpose, and our analyses agree with those from Hokkaido. We propose that this moving probe method of temperature analysis was as successful in the Connecticut River Valley as in Alaska (Benson and Bowling 1973) and warrants additional trial.

DISCUSSION

Nakamura and Magono (1982), Magono et al. (1988), Maki et al. (1986) and Harimaya and Maki (1988) observed inversions that formed over large snow-covered basins, and measured regional minimum temperatures beneath these inversions

using a large array of thermometers and tethersondes. They were able to successfully model temperatures over snow-covered slopes and basins through application of these observations. We have been able to reproduce their observed results over diverse terrain, using a moving probe.

We found that temperature differences with respect to the river plane reference temperature defined at the Piermont (*P*) bridge increased greatly with both distance and elevation along the hilly terrain east of the river. Temperature differences did not increase monotonically with elevation, even when temperature lapse was the dominant inferred vertical structure. Inversion or isothermal structure was intermittently present over small basins, even under general temperature lapse structure. The inversion strength increased as snow covered these small basins, and exceeded 40°C/km when the valley was under general inversion structure. Nakamura and Magono (1982) and Magono et al. (1988) found low inversions formed first over basins, which prevented warmer tropospheric air above from exchanging to the basin surface. This allowed the basin to cool to lesser temperature than surrounding slopes where exchange of warm air from above offset radiational heat loss from the snow surface.

Magono's group worked in a relatively large basin over 100 cm or more of snow. We observed that temperature inversion formed over these small basins along slopes when snow cover was sparse or nil, and that as these small basins became snow covered, the coldest local air temperature was observed in these basins. The lower portion of the larger Connecticut River Basin had lapse structure just above the surface on most days of sparse or thin snow cover even when strong inversion was present 100 m above river elevation. A surface-based inversion was predominantly present in the lower 100 m of the basin when 20 cm or more of snow cover was present.

Baker et al. (1990, 1991) concluded that 5 to 15 cm of snow cover was sufficient to mask agricultural surfaces. They attributed an 8.4°C temperature diminution to snow cover. Our observations indicated that 20 cm of snow was necessary to provide a continuous snow surface that allowed an early and uniform inversion to form over a large basin in more complex topography. This uniformity was evidenced by smoke drift above chimneys indicative of 1-m/s wind by Beaufort analogy, and in agreement with Magono's proposal that diminished surface (1.25-m) temperature accompanies winds of 1 m/s about 9 m above the surface. Inversions and diminished temperatures were found over smaller nearby basins with only a few centimeters of snow cover. We do not have sufficient data to propose a climatically conclusive value of temperature decrease with respect to snow cover, but offer typical differences of –2 to –4°C in small basins relative to surroundings when snow was sparse or nil. Differences of as much as –18°C were found with respect to ridgetops over thicker snow cover. There were several instances when air temperatures diminished more in the valley although cold advection continued at the ridgetops, and many more instances when subzero or diminished temperatures persisted for another morning in the valley after warm advection began on the ridgetops. There is little doubt that snow cover increased the frequency of colder air temperature in this way, but specifying a climatic magnitude would be speculation.

The coldest air was most frequently measured in basin II in the vicinity of *W*. Temperature exploration around the area found temperature inflections or inversions at this elevation along several slopes. In other seasons the top of valley fog was frequently observed just below the elevation of *W*. All of our tabular and graphical analyses show a discontinuity at this level. We propose that an inversion persistently formed at about 100 m above river level, that is, at an elevation of 220 to 250 m. The coincidence of this inversion, with basin II, allowed this basin to cool to lesser temperatures than nearby smaller or larger basins on many nights.

The geometric standard deviation of temperature difference, with respect to the temperature over the river at bridge *P*, was systematically greater at all observing points when inversion was present. Figure 12 shows that under these inversion conditions, small basins have lesser geometric standard deviation of temperature difference than slopes or ridge tops. We interpret this difference in GSD as a measure of the decoupling of basins from exchange with the warmer tropospheric air above. Ridges and slopes remain warmer because of greater air exchange. We propose that the GSD of temperature difference provides an objective analytical index of the relative amount of air exchange that occurs at a selected location. Combining the moving probe temperature measuring technique with analysis of the geometric standard deviation of observed temperatures might provide a relatively rapid and objective insight to exchange and dispersion parameters in a data-sparse winter environment.

CONCLUSIONS OF PART II

Winter morning air temperature differences observed in the Connecticut River Valley are comparable with temperature differences observed near Fairbanks, Alaska, by Benson and Bowling (1973) and in Hokkaido, by Maki and Harimaya (1988). Surface air temperature differences of 5 to 18°C were commonly observed along slopes within 6 km of the river on the same days that less than 2°C temperature difference was observed along a 30-km path along the river. We attribute the largest differences to earlier formation of inversion over small basins and diminished exchange of warmer tropospheric air across these inversions through the night. A discussion of the limited influence of cloudiness was provided in Hogan and Ferrick (1997).

The moving probe method of temperature measurement has produced results consistent with those obtained by a multiple thermometer and tethersonde array in Hokkaido. The inference of temperature inversion or lapse structure in the lower 300 m of air over the Connecticut Valley from surface observations at two elevations has yielded consistent results in describing inversion strength and temperature differences. This technique warrants further field comparison with more sophisticated instrumentation in other terrain settings.

The influence of snow depth and cover on winter morning air temperature has been demonstrated. Lapse rate in the lowest 300 m diminishes and lapse becomes less frequent as snow depth increases. Inversion becomes stronger and more frequent as snow cover increases. A surface-based inversion is infrequent in the lowest 100 m of the river basin when less than 20 cm of snow covers the ground; a surface-based inversion diminishes river level temperatures when more than 20 cm of snow is present.

The diminution of winter morning temperatures in small and large basins results from diminution of exchange of warmer tropospheric air from above during night hours. The geometric standard deviation of temperature difference, with respect to the lowest elevation, is least in basins and greatest on ridges and slopes. We propose that exchange and dispersion parameters may be related to the geometric standard deviation of temperature differences.

PART III: SOME EXAMPLES OF THE INFLUENCE OF CULTURAL AND PHYSICAL FEATURES ON MORNING AIR TEMPERATURE

INTRODUCTION

We established in Parts I and II that the small variation in winter morning air temperature that occurs in the plane coincident with the banks of the Connecticut River is related to stability of the air overlaying the valley and the continuity of snow cover. The vertical temperature structure of the lower 100 m of air overlaying this plane (i.e., that below 230-m elevation) is dependent on the temperature structure of the air above and the continuity of surface snow cover. The frequency of occurrence of inversions, and the magnitude of inversion, increase with the thickness and continuity of snow cover. The mean vertical temperature structure in the study area and analysis of temperatures in basins just above river level, by Hogan and Ferrick (1993), indicate adjacent flats or basins separated by a discontinuity or barrier had comparable temperatures. The techniques of analysis developed in Parts I and II will be used to examine winter morning temperature variation associated with cultural and surface features.

ANALYSIS OF TEMPERATURE DIFFERENCE OCCURRING WITHIN A HAMLET

The magnitude of influence of physical or cultural discontinuities on the morning air temperature can be estimated by comparison with adjacent temperature, in a similar manner to that described by Lowry (1977) and Landsberg (1981). Landsberg (1981) showed that the initial construction beginning the "new town" of Columbia, Maryland, induced a 1°C heat island about the construction site. Our initial work (Hogan and Ferrick 1990) showed "downtown" Hanover, New Hampshire, to be 1 to 2°C warmer than adjacent open fields and recreational areas at the same elevation when the ground was snow covered. We were not able to distinguish or attribute a temperature influence in comparing hamlets of about 30 buildings to surrounding terrain using a single winter of data. Temperatures observed along the green in Lyme, noted by name in the map in Figure 3 are compared with temperatures measured at a similar elevation south of the hamlet in Table 8. The tabulation is again stratified by stability and snow cover, and the temperatures shown are temperature differences, referenced to the river plane at P. The mean temperature difference at u, 29.4 km from P, is shown as a reference. Note that the mean temperature difference at u was shown to approximate the 84th percentile of the frequency distribution of temperature differences between P and u in Part I. The mean temperature difference, exceeded in less than 0.16 of cases in this data set, is shown here for more direct comparison to other heat island analyses.

We have omitted temperature differences measured under lapse in Table 8, as these are consistently small, in agreement with Lowry (1977). There is very little temperature difference under inversion when less than 10 cm of snow is present. When snow exceeds 10 cm, the temperature along the green is 0.5°C greater in the mean than the adjacent temperature, although other nearby temperatures are comparable. This indicates that the wooden buildings along the green, snow removal around the perimeter, and the presence of several idling vehicles along the green induce very little modification of the surface air temperature field. This is probably due to the green, and the surrounding area, providing a relatively uniform snow-covered surface plane much greater in area than that disturbed by buildings and roads. A much more sophisticated experiment would be required to identify a heat island associated with a hamlet of this size. The small, nearly level "flats" are not bounded basins but are embedded on slopes. They are systematically warmer than the Connecticut River reference plane.

Table 8. Temperature differences observed at and near the Lyme Green, with respect to Piermont bridge P temperatures.

		Snow cover category			
Location	*Elevation (m)*	*<3 cm (°C)*	*3–10 cm (°C)*	*10–30 cm (°C)*	*>30 cm (°C)*
Lyme Green	171	(–)0.25	1.05	1.30	4.10
Flat N of t	160	(–)0.55	1.05	0.80	3.50
Flat at t	177	0.15	1.15	1.60	4.70
Flat S of t	177	0.35	0.75	0.90	3.70
River at u	122	1.15	1.40	1.10	2.20

THE "HEAT ISLAND" ASSOCIATED WITH A FREEZING RIVER

The measurement series described in this work began during the winter of 1989–90. The sampling point network was altered several times during that winter, and most data reported have been from the homogeneous measurement sequence used during 1990–93. Temperature measurements in the vicinity of the Piermont bridge, *P*, detected an interesting heat island during early December 1989. The temperature measurement points that defined this heat island are those east of the Piermont bridge on Figure 3.

December 1989 was an unusually cold month in the experiment area, and this month was climatologically one of the coldest Decembers of record in the northeastern United States. The air temperature remained continuously below freezing at the recording site at *L* in Figure 3 throughout the first 30 days of the month. Morning temperatures at *L* were less than –15°C on all but four days. Ice cover formed on the Connecticut River on 4 December. A light snowfall on 6 December allowed visual verification that the ice cover was continuous throughout the study area. A few additional centimeters of snowfall provided thin snow cover through early December.

A significant snowfall of 25–35 cm occurred in the study area on 16 December 1989. We compared (Hogan and Ferrick 1990) the morning air temperatures near the river on the five days prior to this snowfall with those on the days following. The temperatures observed prior to the snowfall, on the bridge *P* and at *h* adjacent to the bridge, were 2°C greater than those observed east of *h*, along basin IV. The temperatures on the bridge, and at *h*, were less than those east of *h* following the snow of 16 December.

We propose that the air over and adjacent to the river was warmed by the heat rejected during river ice growth prior to 16 December, and that the insulating effect of the additional snow halted this warming process. We will examine this hypothesis using the Ashton (1989) ice growth equation.

The heat released due to ice growth is proportional to the change in ice thickness, over a length of river of a given width. Temperature index models provide good estimates of the growth of river ice on a daily scale. The equation for ice growth of Ashton (1989) can be rewritten to include the thermal resistance of snow cover:

$$\frac{dh_i}{dt} = \frac{1}{\rho_i L} \frac{(T_m - T_a)}{\left(\frac{h_i}{k_i} + \frac{1}{H_{ia}} + \frac{h_s}{k_s}\right)} \tag{1}$$

where the *i* subscripts refer to the river ice, the *s* subscripts refer to snow on the ice, and *a* subscripts refer to air above the ice or snow. The terms are defined in *Nomenclature*, and the temperature of the water beneath the ice is fixed at the melting point.

Integrating eq 1 over a finite time increment results in a quadratic equation. The positive root of this quadratic equation can be solved for the final ice thickness h_{if} at the end of the time increment:

$$h_{if} = \left[(h_{ii} + h_r)^2 - \frac{2k_i T_a \Delta t}{\rho_i L} \right]^{1/2} - h_r \tag{2}$$

where

$$h_r = k_i \left(\frac{1}{H_{ia}} + \frac{h_s}{k_s} \right). \tag{3}$$

where h_{ii} is the initial ice thickness and h_r acts as an equivalent ice thickness corresponding to the snow exchanging heat with the air. The squared term in eq 2 is much larger than the quotient term, allowing the braced square root term to be replaced by the first two terms of a converging binomial series. This allows eq 2 to be rewritten as

$$h_{if} = h_{ii} - \frac{k_i T_a \Delta t}{\rho_i L (h_{ii} + h_r)} \tag{4}$$

in which the h_r term acts as an equivalent ice thickness corresponding to the snow exchanging heat with the air.

This can now be used to estimate ice growth on the Connecticut River during early December 1989. Ice cover was observed on the morning of 4 December, indicating ice growth began sometime on 3 December. The mean daily temperature in the plane intersecting the river bank was –14°C during the period 3–15 December. The snowfall of 6 December was quite light, and can be neglected in using eq 2 to calculate a mean ice growth of 3 cm/day (0.03 m/day), yielding 0.36 m (36 cm) of ice for the period.

We now assume that the 25–35 cm of snow which fell on 16 December settled to a rather uniform 20 cm during the following days. Ashton

(1986) shows that the thermal conductivity of a snow layer varies as the bulk density of the snow. Although snow rarely has a uniform density throughout its depth, we can conservatively estimate the bulk density of snow as 200 kg/m^3, with thermal conductivity of 0.12 W/m °C. The mean air temperature of –13°C on 17 December would produce an additional ice growth of 0.002 m, using either eq 2 or 4. The snow cover diminished the rate of river ice growth by a factor of 10, although mean daily air temperature did not appreciably change. This corresponds to a daily latent heat release of 1.4×10^9 J/m of river length prior to the snowfall, and 8.1×10^7 J/m after, when distributed over the mean 150-m width of the river.

The heat transferred to the air above the ice produces a change in enthalpy

$$ {}_1Q_2 = H_2 - H_1 = M\, C_p \left(T_2 - T_1\right) \tag{5} $$

which can be used to estimate the maximum air temperature change which could result from the latent heat of freezing. Values for specific heat and density of air are obtained from Keenan and Kaye (1948) and List (1950). The mass of near surface air M influenced by this heat release must be estimated to evaluate the temperature difference (T_2–T_1).

We showed in Part I that the temperature at h, one river width from the river, was equivalent to the air temperature over the river. We showed in Part II that an inversion was generally present 100 m above river elevation over sparse snow cover, and that an additional inversion was present just above the river elevation when more than 20 cm of snow covered the Connecticut River Valley. The points in basin IV, apparently uninfluenced by the latent heat release, are about five river widths from the river and 20 m elevation above the river. The volume of air influenced by this latent heat source is about 1 km in width and 100 m in depth, that is 10^5 m^3 per meter of river length, over sparse snow cover. This volume diminishes to about 3×10^4 m^3 per meter of river length beneath the lower inversion accompanying more than 20 cm of snow cover. The addition of 1.4×10^9 J/m day could account for a 10°C stagnant air temperature increase, while the addition of 8×10^7 J/m day through the thicker snow cover could account for less than 2°C of rise in a thinner stagnant air layer. The observed mean temperature difference increased 4°C after the snowfall of 16 December.

A more specific analysis compares the day before the snowfall 15 December, T_p = –24.5°C, with a synoptically similar day 18 December, T_p = –25.5°C. The temperatures in basin IV east of the bridge were 2°C less on 15 December, and 2°C greater on 18 December. These differences are 10 times the mean temperature differences attributable to difference in inversion structure with snow cover shown in Part II.

A heat island of approximately 4°C has been found in the vicinity of a freezing, sparsely snow-covered river. This is a greater local heating than associated with some urban heat islands. This heat island was recognized at the time of measurement, and preliminarily reported (Hogan and Ferrick 1990). The magnitude of this observed heat island was reassessed, considering the river plane reference of Part I and the variation of vertical temperature structure in Part II. This has verified that ice growth under sparse snow liberated sufficient heat to increase the temperature of stagnant air 0.4°C/hr. A 4°C temperature increase relative to surroundings was realized under a local inversion that persists about 12 hours.

COLD AIR DRAINS

Observation points m and n, and several other points where brooks or streams enter the lower basin, were initially chosen to examine the frequency of lesser temperature occurring at those places due to cold air drainage. On 13 of the 40 days with lapse conditions, lesser temperatures were observed along the brook mouth at n than at the other points along the river. This was the most frequent occurrence of observed cold air drainage. Interestingly, the coldest air observed along this observation path occurred at m, and at the adjacent Orford Green on 27 January 1994. Cold air drainage may also account for lesser temperatures observed near s on three days. Cold air drainage was not frequently apparent in these observations, but was coincident with increased local temperature difference when it was observed.

The discussion of the variation of vertical temperature structure with respect to snow depth in Part II alluded to the possibility of cold air drainage down the narrow stream valley connecting small basins II and III. The topographic cross sections of this valley and basin II show it to be quite narrow and steep sided, but much less so than the V-shaded valley of Yasuda et al. (1986). This valley is intermittently wooded, among small fields,

Table 9. Frequency distribution of temperature difference along a narrow stream valley.

Location	Elevation (m)	Temperature distribution 84th%	Median	16th%	GSD	Stability
Along *P–O*	122–130	0.5	0.2	<0.1	2.5	L
j	180	0.9	0.4	<0.1	2.3	L
5	181	1.1	0.5	0.1	2.2	L
4	188	1.3	0.6	0.2	2.3	L
3	189	1.2	0.5	0.2	2.3	L
2	204	1.5	0.8	0.3	2.0	L
1	223	1.3	0.6	0.2	2.1	L
W	230	2.4	1.3	0.2	1.8	L
Along *P–O*	122–130	1.0	0.4	0.1	2.5	I
j	180	1.4	0.6	0.2	2.4	I
5	181	1.8	0.9	0.2	2.0	I
4	188	1.7	0.8	0.2	2.0	I
3	189	1.7	0.8	0.2	2.0	I
2	204	2.4	1.0	0.3	2.4	I
1	223	2.9	1.2	0.4	2.4	I
W	230	3.0	1.6	0.5	2.1	I

and originates and terminates in wider flatter basins.

The data used to produce the vertical temperature profiles plotted in Part II, Figures 10 and 11, were combined into two sets dependent on lapse or inversion structure. Frequency analysis of temperature difference at each point from *j* at the western most edge of Basin III to *W* in Basin II are given in Table 9. The five air temperature differences at the near river observing points between the Piermont (*P*) and Orford (*O*) bridges have been combined to provide *P–O* reference differences for the same days in Table 9. The median and GSD of the distribution of these differences provides an insight for comparison of the trend in temperature difference.

The temperature differences observed at 3, 4, 5, and *j*, and the GSD of the differences, are quite comparable under both lapse and inversion conditions. There is systematically less difference at *j* which is closest in proximity to the Piermont bridge (*P*) reference point. It appears that the channeling of flow over snow cover observed by Yasuda did occur here under inversion conditions. The colder air from the basin (II) surrounding *W* appears to extend to 1, along the side of the valley on some mornings with inversion. The standard deviation of temperature difference at points 1 and 2 approach that of *W* under lapse and depart from that of *W* under inversion. We interpret this as evidence that the temperature inflection or inversion that is generally present over the basin (II) surrounding *W* under general lapse continues above 1 and 2, but allows some exchange with layers of warmer air from higher elevation. Intermittent exchange immediately beneath the inversion or inflection may alter the air temperature at points 3, 4, and 5 (basin III) giving the appearance of a superadiabatic lapse rate coincident with the valley as noted in Part II. The stratification over snow-covered basins II and III only allows cold air drainage down the valley when lapse coincides with minimal snow cover.

FREQUENCY OF SUBZERO (<–17.8°C) DAYS IN THE STUDY AREA

Climatological data cited (NOAA 1982) in the Introduction to Part I showed a great variation in the frequency of subzero (<–17.8°C) days in the vicinity of the study area. The results of observations presented to this point have primarily related to temperature differences, with respect to variation in topography and snow cover within the study area. The number of observations of subzero (<–17.8°C) and lesser air temperatures at several observation points are tabulated in Table 10. The tabulation is ordered by decreasing frequency of occurrence of subzero mornings.

These observations represent 158 observation days over three winters when no rain or supercooled fog occurred, and should not be consid-

ered in any way as a climatology. The distances tabulated are relative to W, the point with most frequent subzero days; note that the A is only 0.6 km from c in basin I. There are two lapse days when air of less than –17.8°C was uniquely observed at the apex included in the tabulation. A smaller set of observations along the north–south path including 24 of the days <–17.8°C was observed at P and j. Subzero air was rather uniformly observed along the path on 21 of these days, but no air temperature less than –25°C was observed south of k during 1990–1993. There was one day that air of –18°C was observed along the river south of u, but no air of less than –17.8°C observed elsewhere.

Table 10. Number of observations of lesser temperatures.

Location	*Distance (km)*	*<–17.8°C*	*<–20°C*	*<–25°C*	*<–30°C*	*<–35°C*
W, basin II	0	42	35	16	3	0
c, basin I	2.0	40	31	8	3	1
z (lower case)	0.2	37	28	7	1	0
g	3.1	35	25	5	1	0
Bridge *P*	5.6	34	25	6	1	0
j	3.3	33	26	6	1	0
A, apex	2.6	29	20	2	0	0

Table 11. Frequency of winter morning temperatures at Piermont bridge (*P*).

	<3 cm	*3–10 cm*	*10–20 cm*	*>20 cm*
Number of observations	43	31	34	45
Number less than –17.8°C	6	3	8	14
Number less than –20.0°C	2	1	5	14
Number less than –25.0°C	0	0	0	5
Number less than –30.0°C	0	0	0	1
Median temperature, °C	–7.6	–8.0	–11.4	–11.8

We have shown in Part II that frequency of lapse and inversion structure systematically varied with the depth and continuity of snow cover. We also showed that the frequency of greater local temperature differences, and the mean value of local temperature differences, increased as snow cover increased in depth and uniformity. All of these systematic differences were temperature difference with respect to the temperature measured at P, the northernmost river level measuring point along the observation path. The temperatures observed at P are stratified with respect to snow cover category in Table 11. The entire set of temperatures observed at P are near normally distributed, but stratified segments depart from normality. We have used the 10- to 20-cm class described in Part 1, and combined the 10- to 20-cm class with the >30-cm class to provide four snow cover categories with approximately equal numbers of observations. The table indicates that during this series of observations lesser temperatures, and more frequent subzero (<–17.8°C) temperatures, accompanied thicker and more uniform snow cover at P.

Considering these combined arguments indicates that P more often experiences lesser air temperatures, with respect to its surroundings, as snow cover thickens and becomes more uniform. This lesser temperature is associated with more frequent formation of inversion structure along the plane intersecting the river bank as snow thickness increases.

DISCUSSION

The introduction to Part I showed that climatic time scale differences of a factor of two in the annual number of subzero (<–17.8°C) days occurs over distances of 30 km in northern New York and New England. Part I of this work showed that temperature differences along a 30- to 50-km homogeneous plane coincident with the bank of the Connecticut River were on the order of 1°C or less on a majority of 158 observing days over three winters, including 40 subzero days. Quasi-simultaneously, temperature differences of as much as 18°C were observed at points within 6 km of the river. The number of mornings with subzero temperatures differed by one-third at two points separated by less than 0.6 km in distance and 60 m in elevation.

The most frequent occurrences of subzero morning temperature are at W, which is also the point with the least mean and median morning air temperature. The least frequent occurrences of

subzero air temperature were at A, coincident with the greatest mean morning temperature in the data set, and the greatest GSD of temperature difference with respect to the reference temperature at river level.

The reference temperature at P systematically diminished as snow thickness increased and snow cover became more uniform. The diminution of temperature was evidenced by a diminished mean morning temperature, a diminished median temperature, and more frequent subzero morning temperature coincident with increasing snow cover.

The morning air temperature observed in a hamlet could not be systematically differentiated from those measured on nearby flats of similar elevation. Temperatures observed along a narrow creek valley, and those observed where side streams entered the river basin, were insignificantly different from adjacent temperatures at similar elevation on mornings with inversion. Cold air drains were observed on some mornings when lapse occurred concurrently with sparse snow cover, and cold air drains occasionally produced asystematic lesser temperatures.

Hill and ridge tops, adjacent slopes, and small basins along slopes consistently exhibited the greatest morning air temperature differences studied. Greater snow depth, and its coincident greater uniformity of cover, increased these differences.

The experiments and models of Magono et al. (1982) and Maki et al. (1986), in conjunction with the observations in a different geographic setting with more variable snow cover in this work, may aid interpretation and extrapolation of meteorological records over snow-covered, data-sparse, and complex topography. The observations and analyses of Parts I, II, and III agree with the Hokkaido model theory, showing slopes to be relatively warmer, and basins relatively cooler on snow-covered winter mornings. The dynamic moving probe method of temperature observation showed slope, basin size, and snow cover to be the dominant features determining winter morning air temperature in the vicinity of the Connecticut River. Temperature differences associated with hamlets, barriers, and surface vegetation were minor when compared. Small basins influenced local temperature and temperature structure even when strong cold advection occurred over sparse snow cover. The larger basin along the Connecticut River did not frequently produce inversion structure in the lowest 100 m of air when less than 20 cm of snow covered the ground. The Connecticut River Basin consistently produced inversion structure in the lowest 100 m of air when more than 30 cm of snow cover was present, in agreement with Magono's findings over deep snow in the Moshiri Basin.

CONCLUSIONS OF PART III

The number of subzero (<–17.8°C) mornings varied by one-third, among observation points separated by a distance of less than 6 km, in our 158-day experiment. The number of mornings less than –25°C had even greater variation. This difference is attributed to earlier formation and later persistence of near surface inversions over small basins, and more persistent exchange of warmer air from aloft to slopes and ridges, in agreement with the theory and observations of Magono et al. (1982) and Maki et al. (1986). We were not able to isolate significant temperature difference attributable to hamlets or vegetation in this winter study.

Our hypothesis that a river plane provides a temperature reference in winter, and that a moving probe senses multiple surface temperatures that can be used to synthesize the vertical temperature structure in hilly terrain, has provided consistent and useful results in this case. This hypothesis seems worthy of additional examination in other terrain. The combination of moving probe observation, with frequency analysis of temperature difference, provides the geometric standard deviation of local temperature difference. The magnitude of the GSD appears related to air exchange over snow-covered ground.

We propose that the principles of constructing vertical temperature structure using a moving probe, and the analysis of the frequency distribution of temperature differences as an estimate of local exchange, can be applied to prepare local dispersion estimates in data-sparse operational settings.

We showed through analysis of the frequency distribution and mean value of temperature differences in Parts I and II that a relatively repeatable pattern of temperature difference, with respect to elevation and snow cover, is present above the river surface and east of the Connecticut River on winter mornings. Collections of observations, obtained at more than one observation point at the same (±5-m) elevation, were used in preparation of these figures. Several of these collections grouped points from fields and woods,

across forest edges, and through a narrow valley that intersected the most frequent inversion level. The frequency distribution of temperature differences at the individual observing points along the shortest route from the river plane to the level of the most frequent inversion base are given in Table 9 for both lapse and inversion conditions. The route is shown in Figure 3, and the coordinating symbols may be used to locate or compare this data with respect to Parts I and II.

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REPORT DOCUMENTATION PAGE

Form Approved
OMB No. 0704-0188

1. AGENCY USE ONLY (Leave blank)	2. REPORT DATE	3. REPORT TYPE AND DATES COVERED
	December 1997	

4. TITLE AND SUBTITLE	5. FUNDING NUMBERS
Local Variation in Winter Morning Air Temperature	PR: 4A161102BT25 WP: 304 WU: EC-B03
6. AUTHORS Austin W. Hogan and Michael G. Ferrick	
7. PERFORMING ORGANIZATION NAME(S) AND ADDRESS(ES) U.S. Army Cold Regions Research and Engineering Laboratory 72 Lyme Road Hanover, New Hampshire 03755-1290	8. PERFORMING ORGANIZATION REPORT NUMBER CRREL Report 97-9
9. SPONSORING/MONITORING AGENCY NAME(S) AND ADDRESS(ES) Office of the Chief of Engineers Washington, DC 02314-1000	10. SPONSORING/MONITORING AGENCY REPORT NUMBER

11. SUPPLEMENTARY NOTES

For conversion of SI units to non-SI units of measurement, consult ASTM Standard E380-93, *Standard Practice for Use of the International System of Units,* published by the American Society for Testing and Materials, 1916 Race St., Philadelphia, Pa. 19103.

12a. DISTRIBUTION/AVAILABILITY STATEMENT	12b. DISTRIBUTION CODE
Approved for public release; distribution is unlimited. Available from NTIS, Springfield, Virginia 22161	

13. ABSTRACT (*Maximum 200 words*)

Results of temperature measurements, which may be applied to inference of winter temperatures in data-sparse areas, are presented. The morning air temperatures during three winters were measured at 80 places in a 10-× 30-km area along the Connecticut River. NOAA climatologies show this region to have complex spatial variation in mean minimum temperature. Frequency analysis techniques were applied to evaluate the differences in daily local temperature. Temperature lapse or temperature inversion in the study area was inferred from the difference of surface temperature measurements 100 and 300 m above river level. The frequency of inferred temperature lapse and the inferred lapse rate diminished as snow cover increased. The frequency of inferred temperature inversion and inversion strength increased as snow cover increased. When more than 20 cm of snow covered the ground, an additional surface inversion was frequent in the layer less than 100 m above river level, and two-thirds of river level temperatures less than −20°C occurred concurrent with these conditions. The daily temperature differences at the individual points, with respect to a defined point, were lognormally distributed. The magnitude and geometric standard deviation of temperature differences throughout the study area were larger on mornings when inversion was inferred. With respect to topography, temperature differences and the geometric standard deviation of temperature differences were smaller along flats or among basins than along or atop slopes on mornings when inversion was inferred. It is proposed that some meteorologically prudent inferences of surface temperature and near-surface temperature lapse or temperature inversion can be made for similar data-sparse areas.

14. SUBJECT TERMS			15. NUMBER OF PAGES 50
Air temperature Connecticut River	Heat islands Meteorological inversions	Winter	16. PRICE CODE

17. SECURITY CLASSIFICATION OF REPORT	18. SECURITY CLASSIFICATION OF THIS PAGE	19. SECURITY CLASSIFICATION OF ABSTRACT	20. LIMITATION OF ABSTRACT
UNCLASSIFIED	UNCLASSIFIED	UNCLASSIFIED	UL